Retrait de la subduction et exhumation des unités de haute pression dans les Alpes

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Aegean tectonics:
strain localisation, slab tearing and trench retreat

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Abstract : The geology of the Aegean-Anatolia region has fostered opposing views on the rheology at lithospheric scale: the continental lithosphere is either resistant and made of intrinsically strain-localising materials leading to the formation of large-scale strike-slip faults or an overall weak medium that is able to distribute the strain over large domains. We review the geodynamic evolution of this region and discuss the processes leading to strain localisation over geological times. From Late Eocene to Present, crustal deformation in the Aegean backarc has progressively evolved toward a stronger degree of localisation during slab retreat. In the Late Eocene extension started with the formation of the Rhodope Metamorphic Core Complex and in the northern Aegean and in the Oligocene and Miocene, extension was accommodated by several crustal-scale detachments that led to the formation of a first series of core complexes (MCCs) in the Cyclades and the northern Menderes massif. Extension then localised in Western Turkey, the Corinth Rift and the external Hellenic arc after Messinian times, while the North Anatolian Fault penetrated the Aegean Sea. Throughout this evolution the direction and style of extension with shallow north-dipping shear zones and brittle faults have not changed significantly except in terms of localisation. The respective contributions of progressive slab retreat and tearing, basal drag, extrusion tectonics and tectonic inheritance are discussed and we favour a model (1) where slab retreat is the main driving engine, (2) successive slab tearing episodes are the main cause of this stepwise strain localisation and (3) the inherited heterogeneity of the crust is a major factor for localising detachments. The continental crust has inherited a strong heterogeneity from its earlier tectonic history and crustal-scale contacts such as major thrust planes act as weak zones or as zones of contrast of resistance and viscosity that can localise later deformation during exhumation and backarc extension. The dynamics of slabs at depth and the asthenospheric flow due to slab retreat also have a major influence on the localisation of deformation in the upper plate. Successive slab ruptures along the strike of the Hellenides and Taurides from the Middle Miocene to the Late Miocene have progressively isolated a narrow stripe of lithosphere, still attached to the African lithosphere at the longitude of Crete. The formation of the North Anatolian Fault and its propagation within the Central Hellenic Shear Zone is partly a consequence of this evolution. Once the connection of the NAF with the trench through the CHSZ has been made the extrusion of Anatolia seems to have reoriented the flow of the upper crust and the strain in the lower crust. The extrusion of Anatolia and the Aegean extension are thus partly driven from below (asthenospheric flow) and from above (extrusion of a lid of rigid crust).

1. Introduction

1.1 Different visions of the rheology of the continental lithosphere

There is still no consensus on the mechanical behaviour of the continental lithosphere. Experimental observations on rock mechanics led to the formulation of rheological yield-stress envelopes that explain reasonably well the brittle-ductile layering and long-wavelength phenomena such as lithospheric flexure (Goetze and Evans, 1979; Molnar, 1992; Kohlstedt et al., 1995; Jackson, 2002; Armijo et al., 2003; Watts and Burov, 2003; Handy and Brun, 2004; Burov and Watts, 2006). However, lithospheric deformation and rheology appear more complex as it becomes more intense and localised (Handy et al., 2007; Bürgmann and Dresen, 2008; Gueydan et al., 2003,2004; Précigout and Gueydan, 2009; Burov, 2011). So, two end-member models have been discussed (Goetze and Evans, 1979; Molnar, 1992; Kohlstedt et al., 1995; Jackson, 2002; Armijo et al., 2003; Watts and Burov, 2003; Handy and Brun, 2004; Burov and Watts, 2006). They oppose by a more or less important propensity to localise deformation, some explaining the propagation of strike-slip faults over large distances (Tapponnier et al., 1982; Armijo et al., 2003), others, more “ductile”, explaining continental extension or shortening over large areas (England and Houseman, 1986; Wernicke, 1992). It is also argued that plate boundaries are characterized by some very specific rheological properties (Bürgmann and Dresen, 2008). This has resulted in a specific “banana split” yield-stress model that considers the weakness of major crustal fault zones caused by various strain weakening and rheological feedback processes.

The observations of localised large-scale strike-slip faults in the continental lithosphere have raised the problem of continental extrusion (or escape tectonics). After the large-scale Asian strike-slip faults were described north of the Indian indenter and the rigid-plastic indentation model published (Molnar and Tapponnier, 1975; 1978; Tapponnier and Molnar, 1976; 1977), the conceptual jump to the extrusion model (Tapponnier et al., 1982; 1986) has not been accepted by all researchers. Two trends have been independently followed, extrusion and localising rheology on one hand (Tapponnier et al., 1982; 1986), or distributed deformation of Asia and the Tibetan plateau on the other hand (England and Houseman, 1986; Dewey et al., 1988; England and Molnar, 1990; 1997; Royden et al., 1997). This discussion results from different visions of the rheology of the continental lithosphere in the international community.

1.2 The Aegean domain: a natural laboratory for the rheology of the continental lithosphere

This debate has lasted for more than 30 years and it is nowadays focused on the Aegean region where different sets of data are available, besides a detailed knowledge of the geological evolution at crustal scale (Aubouin, 1959; Brunn et al., 1976; Jacobshagen et al., 1978; Le Pichon and Angelier, 1979; 1981b; 1981a; Angelier et al., 1982; Bonneau and Kienast, 1982; Bonneau, 1984; Sotiropoulos et al., 2003; Papanikolaou et al., 2004; van Hinsbergen et al., 2005a; Jolivet and Brun, 2010; Ring et al., 2010): a precise present-day velocity field measured by satellite geodesy (McClusky et al., 2000; Le Pichon and Kreemer, 2010; Reilinger et al., 2010), a good description of Quaternary geology and morphology (Mercier et al., 1979; Mercier, 1981; Jackson et al., 1982; Jackson and McKenzie, 1984; Armijo et al., 1996; 2003; Goldworthy and Jackson, 2001; Rohais et al., 2007a), an Oligo-Miocene strain field, since ~30-35Ma (Lister et al., 1984; Sokoutis et al., 1993; Gautier and Brun, 1994b; 1994a; Brun and Sokoutis, 2007; Grasemann and Petrakakis, 2007; Iglseder et al., 2009; Jolivet and Brun, 2010), or since 23 Ma (Ring et al., 2010) recovered from the middle to deep crustal exhumed rocks, as well as recent advances on mantle structure from seismic tomography and SKS anisotropy below the Aegean and Anatolia (Wortel and Spakman, 1992; 2000; Hatzfeld et al., 2001; Piromallo and Morelli, 2003; Faccenna et al., 2003; 2006; Spakman and Wortel, 2004; Govers and Wortel, 2005; Jolivet et al., 2009; Suckale et al., 2009; Endrun et al., 2011; Evangelidis et al., 2011; Salaün et al., 2012).

Recent evidence for propagation of the North Anatolian Fault over a distance of one thousand kilometres suggests a long-term elastic behaviour of the lithospheric mantle (Armijo et al., 1999; 2003). Sengör et al. (2005) instead favour a more progressive localisation of the NAF starting from a wide dextral shear zone as soon as 11-13 Ma. Similarly, extensional deformation over large regions in the Aegean Sea (Lister et al., 1984; Jolivet et al., 2004a; Tirel et al., 2004a; 2008), and the exhumation of high temperature metamorphic rocks suggest that the middle and lower crusts are weak.

1.3 The rheological behaviours of the Aegean lithosphere

The Anatolia-Aegean Sea region (figure 1A) has been the focus of an international effort during the last 20 years and the conjunction of geological research and the acquisition of GPS data put this region at the centre of the current debate on the mechanics of the continental lithosphere. Previous works in this region reflect the above mentioned debate, from those describing the long-term deformation of the Aegean Sea and the exhumation of extensional metamorphic cores since ~30 Ma (Jolivet and Brun, 2010; Ring et al., 2010), and those describing the interaction between recent extension in the Aegean since ~15 Ma and strike-slip faulting along the North Anatolian Fault since it reached the Aegean domain ~5-6 Ma ago (Armijo et al., 1996; 1999; 2003). These various deformations affect however the same lithosphere and the time periods overlap.

In more details, three differing conceptions of the bulk rheology of the lithosphere in the Aegean region are currently debated. Correspondingly, in those views the continental lithosphere is either (1) viscous and drives from below the motion of upper crustal blocks (models based on continuum mechanics, e.g., (England and McKenzie, 1982; Taymaz et al., 1991; Jackson, 1994; Jolivet et al., 1994a; 2004a; Gautier et al., 1999;), or (2) kinematically rigid, i.e. strain-less, where all displacements occur at the boundaries of rigid-blocks and resulting in a kinematic description close to plate tectonics, e.g., (Le Pichon et al., 1995; McClusky et al., 2000; Nyst and Thatcher, 2004), or (3) is strain-displacement partitioned involving frictional slip on faults and elastic deformation between faults (based on the concepts of fracture mechanics, e.g., (Armijo et al., 2003; Hubert-Ferrari et al., 2003; Flerit et al., 2004). End-member models based on block or continuum deformation can fit the GPS data, but fail to account for some essential aspects of the Aegean tectonics and NAF-Aegean interaction over the geological time scale.

So far, available models have most of the time neglected two important factors, the inherited mechanical anisotropy of the continental crust and the role played by the asthenospheric flow in driving crustal deformation. Although they do not give a detailed picture of lithospheric structure, available tomographic models suggest that the lithosphere is very thin below the Aegean region (Spakman et al., 1993; Piromallo and Morelli, 2003; Salaün et al., 2012) as in most active backarc regions. In such conditions asthenospheric flow has a stronger impact on crustal deformation. As most continental regions, the Aegean has a complex tectonic history that has produced a strong heterogeneity in the crust (i.e. large thrusts, large exhumation shear zones or extensional detachments). This strong mechanical heterogeneity can also play an important role in the localisation process. The mechanical consequences of such heterogeneous features have not been taken into account in general and the Aegean lithosphere is often treated as a stack of homogeneous layers that are either able or unable to propagate localised shear zone or brittle faults over large distances. Numerical simulations have however shown that the presence of weak heterogeneities, in particular, low-strength heterogeneities resulting from nappe stacking, strongly controls the localisation of strain at crustal scale and that the mechanical stratification inherited from previous episodes is also important (Le Pourhiet et al., 2004; Mattioni et al., 2006; Huet et al., 2011a; 2011b; Lecomte et al., 2011), like strain softening along faults influences the symmetric or asymmetric character of extension at lithospheric scale (Huismans and Beaumont, 2002).

Whether crustal deformation in continental environments is driven from above (stresses transmitted horizontally through the crust) or from below (basal shear due to asthenospheric flow) depends upon the tectonic situation and the thermal state of the lithosphere (Molnar, 1988; Bokelmann, 2002; Tikoff et al., 2004). Most models for the Aegean do not consider a contribution of the asthenospheric flow to crustal deformation through a viscous coupling, as may be suggested by the parallelism between stretching directions in the deep crust sampled in exhumed metamorphic core complexes and flow lines in the mantle derived from the analysis of SKS seismic anisotropy (Hatzfeld et al., 2001; Kreemer et al., 2004; Jolivet et al., 2009; Le Pichon and Kreemer, 2010). Indeed, the correlation between seismic anisotropy and crustal deformation is difficult to explain on the basis of asthenosphere-driven flow alone. In case of viscous coupling between the asthenosphere and lithosphere, the associated shear stress would be on the order of 0.001-0.1 MPa (assuming asthenospheric viscosities on the order of 10^{19} Pa s and strain rates on the order of $10^{-16}s^{-1}$ - $10^{-14}s^{-1}$). These values are at least two orders of magnitude below the weakest estimate of lower crustal strength (Bürgmann and Dresen, 2008). Yet, they can result in tectonically significant forces (~ 10^{11} N) when integrated over several-thousand km^2. Asthenosphere-lithosphere coupling may also occur indirectly through gravitational forces produced by dynamic topography, or thermo-mechanical erosion of the crust and lithosphere by hot asthenospheric flow.

After a summary of 30 years of research in this region, we present a synthesis of the tectonic and metamorphic evolution of the Aegean since the Eocene, with an emphasis on the distribution or localisation of strain. We then summarize this evolution of a strongly anisotropic lithosphere in terms of progressive strain localisation within a continuum of N-S to NE-SW backarc extension and discuss the possible drivers, the Arabia-Eurasia collision, slab tearing and sub-lithospheric mantle flow.

It must be first noticed that the use of the terms “backarc extension” can be partly misunderstood in the Aegean because the present-day arc lies right in the middle of the extended domain. It is quite clear also that, in the geological past, the same situation has already occurred. To overcome this difficulty, Ring and Layer (2003) use instead “intra/back arc” extension. The best solution would probably be to use “upper plate extension”. However most papers use the classical terminology “backarc” extension and we will stick to this solution in the present paper.

2. Evolution of ideas since the seventies

2.1 Large-scale geodynamics, slab retreat vs *extrusion tectonics*

The Aegean region has been the subject of numerous studies since the 70’s and the idea that extension results from slab retreat was first proposed as early as 1979 (Le Pichon and Angelier, 1979; 1981a). At that time some first order features were still unknown. For example the Aegean metamorphic core complexes had not yet been discovered (not before the seminal paper of Lister et al. (1984)), and the Rhodope was still a Precambrian core (Zagorchev, 1998). The front of subduction was misplaced in the Hellenic trench instead of south of the Mediterranean Ridge, that was not yet fully interpreted as an accretionary wedge, although the compressional structures have been recognized quite early (Finetti, 1976). Le Pichon and Angelier (1979) proposed that backarc extension in the Aegean Sea resulted from a combination of gravitational forces (1) in the thickened crust and (2) in the dense subducting slab that led to slab retreat. Alternative models considered crustal collapse as the driver of internal extension and peripheral thrusting (Berckhemer, 1977 ; Horváth et al., 1981; Horvath and Berkhemer, 1982) or extrusion tectonics as the cause of formation of the main marginal seas (Tapponnier, 1977). Gravitational spreading has then been explored further by means of analogue modelling (Hatzfeld et al., 1997; Gautier et al., 1999) and extrusion tectonics remained the main driver of the westward motion of Anatolia in the recent models of Armijo et al. (1999; 2002; Flerit et al., 2004).

Studies of earthquakes and active deformation have shown the kinematic relations between dextral shear along the North Anatolian Fault and the Aegean extension (McKenzie, 1972; 1978). Several improvements were then achieved by more precise seismotectonic studies and analysis of the focal mechanisms of earthquakes, constrained by waveform modelling and polarities of first motions (Taymaz et al., 1991; Barka, 1992; Jackson, 1994; Armijo et al., 1996; 1999). With the addition of paleomagnetic studies that suggested fast rotations of crustal blocks during the Miocene (Kissel and Laj, 1988), several models were proposed for recent and active deformation. With their broken-slats model, Taymaz et al.

(1991) assumed that the westward motion of Anatolia induced E-W shortening in the Aegean region because the rotation of the western part of the region was not fast enough, thus inducing N-S extension, allowed by roll-back of the Hellenic subduction. The model proposed by Armijo et al. (1999) is closer to the extrusion model and the Aegean grabens form as tension gashes in the propagating end of the NAF. Note that here, in both approaches, slab roll-back creates the space for extension but is not the source of extension.

A considerable advance was made when space geodetic data became available (Billiris et al., 1991; Kahle et al., 1995; Le Pichon et al., 1995; Davies et al., 1997; McClusky et al., 2000). Confirmed by more recent studies (Nyst and Thatcher, 2004; Aktug et al., 2009; Floyd et al., 2010; Le Pichon and Kreemer, 2010; Reilinger et al., 2010) these data show that the southern part of the Aegean domain moves southward considerably faster than Anatolia moves westward, implying that the driving engine has to be looked for to the south (Doglioni et al., 2002), slab roll-back being the most likely candidate. More detailed palaeomagnetic studies have also constrained more precisely the timing of blocks rotations (Dimitriadis et al., 1998; Duermeijer et al., 1998; 2000; Kissel et al., 2002; Mattei et al., 2004; van Hinsbergen et al., 2005b; 2006; 2010).

2.2 Geology and slab structure in the Aegean Sea

The Aegean region, located in the overriding plate of the Hellenic subduction zone, has been subjected to extensional tectonics since the late Eocene-Early Oligocene (~35 Ma) (Jolivet and Faccenna, 2000; Jolivet and Brun, 2010); see a discussion on the timing of extension below. Earlier extension may have occurred to the North in the Rhodope massif since some 45 Ma at a slower pace (Brun and Sokoutis, 2007). The Hellenides (Figures 1A & 2) formed from the Late Jurassic to the Present above the Hellenic subduction (Bonneau and Kienast, 1982; Jolivet et al., 2003; Papanikolaou et al., 2004; van Hinsbergen et al., 2005a; Brun and Faccenna, 2008; Jolivet and Brun, 2010; Ring et al., 2010; Royden and Papanikolaou, 2011; Philippon et al., 2012). They result from the offscrapping of crustal units from the Pelagonian in the north and then the Pindos ocean and Apulian block further south that were subducted below Eurasia after the closure of the Vardar ocean in the Late Cretaceous. The shortening of the Pindos and Apulian blocks led to the formation of a series of large scale nappes, all emplaced with a south or southwest vergence of the thrust front, from the Eocene to the present (Godfriaux, 1962; Jacobshagen et al., 1978; Sotiropoulos et al., 2003; van Hinsbergen et al., 2005c)

Moreover, in the late 70's, the geology of the Hellenides and of the Aegean Sea (Brunn, 1956; Aubouin, 1959; Godfriaux, 1962; McKenzie, 1972; Brunn et al., 1976; Jacobshagen et al., 1978; McKenzie, 1978; Mercier et al., 1979) was described with a luxury of details, but some major aspects were still missing such as the massive presence of recent high-pressure and low-temperature (HP-LT) rocks such as blueschists and eclogites (Blake et al., 1981; Bonneau and Kienast, 1982) and the large scale detachments and core complexes that we know today. The geology of Turkey was also quite early explained in the framework of plate tectonics (Dewey and Sengör, 1979; Sengör and Yilmaz, 1981). Most models, for instance the model of Bonneau (1982; Bonneau and Kienast, 1982), used several subduction zones that jumped southward through time, instead of a single migrating subduction used in most models nowadays. The inception of subduction and extension was supposed to be a late feature as young as 13 Ma (Le Pichon and Angelier, 1979) or even 5 Ma (McKenzie, 1978). A breakthrough occurred when with seismic tomographic models that showed a slab much longer than expected (1500 km) and thus suggested a much longer history of subduction and extension with a single slab (Spakman et al., 1988; 1993; Bijwaard et al., 1998), an idea that has been widely developed from then on (Wortel and Spakman, 1992; 2000; Faccenna et al., 2003; Jolivet et al., 2003; van Hinsbergen et al., 2005a; Brun and Faccenna, 2008; Jolivet and Brun, 2010; Ring et al., 2010).

After the closure of the Vardar ocean in the Late Cretaceous, the Apulian domain, including the Pindos ocean, was accreted to the continental margin of Eurasia, building an orogenic wedge with a migration of thrusts from NE to SW. From north to south and structurally top to bottom (figure 1A) the paleogeographic domains involved in the orogen are (see Bonneau, 1984, Jolivet et al. 2004, Papanikolaou et al. 2004, van Hinsbergen et al., 2005 and Brun and Sokoutis, 2007 and references therein for details): (1) The Pelagonian domain, a continental basement and its Paleozoic and Mesozoic cover topped with an ophiolite nappe obducted at the end of the Jurassic. It crops out also in the Cyclades as small klippes or extensional allochtons above the main detachments; it is known as the Upper Cycladic Nappe. This nappe also includes remnants of the Vardar oceanic units obducted in the Late Cretaceous. (2) The Pindos nappe, an oceanic domain or a thinned continental crust with mafic intrusions topped by a pelagic cover (Upper Triassic to Eocene or locally Lower Oligocene). It is found in the Hellenides as the Pindos nappe proper, where it is slightly metamorphosed and also in the Cyclades as a metamorphic equivalent, the Cycladic Blueschists (CBS), reaching locally the eclogite facies. The Ambelakia unit in the Mount Olympos and Ossa regions also belongs to the same realm. The main point in favour of this correlation between the CBS and the Pindos Nappe is their similar tectonic position between the Gavrovo-Tripolitza nappe below and the Pelagonian above (Blake et al., 1984; Bonneau, 1984). The Pindos Ocean was progressively incorporated in the subduction zone and the accretionary wedge and a part of it (to the west) remained unsubducted until the early Oligocene while the CBS were already deeply buried. (3) The Gavrovo-Tripolitza nappe, a carbonate platform (Triassic to Lower Oligocene), most often without basement. It is also found in the internal zones in the core of tectonics windows (i.e. Olympos, Ossa, Almyropotamos, Kerketas Nappe in Samos) (Godfriaux, 1962; Ferrière, 1982; Godfriaux and Ricou, 1991; Ring et al., 1999b; Shaked et al., 2000). (4) The Phyllite-Quarzite unit rests in tectonic contact below the Gavrovo-Tripolitza nappe although it was probably the stratigraphic base of this unit. It is made of a sequence of detrital rocks, including volcanites, limestones as well as basement lenses with ages ranging from the Carboniferous to the Mid Triassic. (5) The Ionian (or Plattenkalk) and the pre-Apulian platform are the two outermost nappes with a Mesozoic carbonate sequence covered with Oligocene-Miocene turbidites.

The Aegean domain, since the Oligo-Miocene, in a geodynamic sense, also encompasses a part of western Anatolia. The Menderes massif has indeed recorded tectonic events that are typically Aegean and it is thus useful to review the evolution of ideas on this region as well. Moreover, the crust is thicker in the Menderes massif and the pre-extension structures are thus better preserved than in the Cyclades.

An additional major step forward was made in 1984 with the first description of a Cordilleran-type metamorphic core complex in the Cyclades on the example of Naxos (Lister et al., 1984). This discovery fostered a renewal of geological studies in the Cyclades (see below). The progressive description of the Cycladic metamorphic core complexes led to several two-stage models, all involving a first period of crustal thickening in high pressure and low temperature (H*P*-L*T*) conditions in the Eocene followed by a period of crustal thinning in H*T*-L*P* conditions in the Oligocene and Miocene while H*P*-L*T* conditions were reached in the external zones (Crete and Peloponnese). In more details, nappe stacking and crustal thickening occured in H*P*-L*T* metamorphic conditions near the front of subduction (accretionary wedge) while crustal thinning was active in the backarc region. These paired belts of compression and extension migrated with time toward the south and the thickened domain was reworked by extensional structures (Jolivet et al., 1994b; Jolivet and Patriat, 1999). The southward younging of H*P*-L*T* metamorphic rocks should be paralleled with the outward migration of thrust fronts and migration of the volcanic arc (Fytikas et al., 1984) at a rate of ~3 cm/yr (Jolivet and Brun, 2010) (figure 3). Reconstructions of the evolution of the

Hellenides and of the backarc domain show the migration of paired frontal compression and backarc extension during slab retreat. A double gradient of finite stretching is observed from the Hellenides in the west and Anatolia in the east toward the centre of the Aegean Sea. Differences in P/T ratio between the Cycladic blueschists and the Cretan blueschists likely reflect this progessive soutward migration (Jolivet et al., 2003; Gueydan et al. 2009).

2.3 The Menderes Massif and western Anatolia

The Aegean extensional domain encompasses a part of western Anatolia, where the relief was strongly shaped by the late formation of core complexes and grabens. In Anatolia, a similar orogenic wedge to the Aegean one can be described, although correlating both regions has been a debated issue for several decades (Brunn et al., 1976; Ricou et al., 1986; Dercourt et al., 1993; Ring et al., 1999a; Jolivet et al., 2004b; Barrier and Vrielinck, 2008).

Remnants of the Tethys Ocean and H*P*-L*T* metamorphic units (Figure 1A) extend eastward within Anatolia (Okay, 1989; Okay and Tüysüz, 1999; Okay et al., 2001; Pourteau et al., 2010). The suture extends into Anatolia as the Izmir-Ankara suture, which west of Ankara branches into the Inner-Tauride suture (Okay and Tüysüz, 1999; Schmid et al., 2011). These separate to the north, continental domains (the Pontides and the Central Anatolia Crystalline Complex) that hosted a magmatic arc between the Late Cretaceous and the early Cenozoic, and to the south a microcontinental fragment, named the Anatolide-Tauride Block (Okay and Tüysüz, 1999) (Figure 1A), which was overthrust by Neotethyan ophiolites and experienced subduction metamorphism along its northern edge. This microcontinent, which comprises a Mesozoic carbonate platform and its Precambrian-Palaeozoic substratum (Bozkurt and Oberhänsli, 2001), was subdivided into several regionally-metamorphosed units to the north, named the Anatolides, and their non-metamorphosed equivalent to the south, named the Taurides.

The Tavşanlı Zone, which is composed of distal sediments and volcanic rocks, experienced subduction metamorphism (e.g., Okay, 1984; Okay, 2002; Çetinkaplan et al., 2008) in the Late Cretaceous (88-80 Ma) (Okay et al., 1998; Sherlock et al., 1999; Seaton et al., 2009), while the Afyon Zone and its far-transported equivalent, the Ören Unit, have recorded low-grade high-pressure metamorphism (Oberhänsli et al., 2001; Rimmelé et al., 2003a; Rimmelé et al., 2006) of Palaeocene age (65-59 Ma) (Pourteau, 2011).

The Menderes Massif (Paréjas, 1940) crops out as a large tectonic window within the Anatolides. It displays a series of basement units, locally with Proterozoic metamorphism, deformed and metamorphosed during the Eocene and later (Hetzel et al., 1995a; Ring et al., 1999a; Bozkurt and Satir, 2000; Bozkurt, 2001; Bozkurt and Oberhänsli, 2001; Lips et al., 2001). The Menderes massif crops out south of the Bornova Flysch Zone (Okay et al., 1996) and the Izmir-Ankara suture zone (Sengör et al., 1984; Okay, 2001; Okay et al., 2001) and north of the Lycian ophiolite (Graciansky, 1966; Bernoulli et al., 1974; Gutnic et al., 1979; Collins and Robertson, 1998). It underthrusts the Cycladic Blueschists to the west (Candan et al., 1997; Oberhänsli et al., 1998; Ring et al., 1999a) and the Afyon Zone to the east. Its definition has evolved through time.

The general stratigraphy of the Menderes massif has been established for a long time (Phillipson, 1918; Erentöz, 1956)) and all other studies used the same general framework (Graciansky, 1966; Dürr, 1975; Sengör et al., 1984; Dora et al., 1990). It consists of a gneissic core, a schist cover and a marble cover. The lack of obvious metamorphic break between the marble cover and overlying Ören Unit (previously considered as part of the Lycian Nappes) led to some confusion on the southern limit of the Menderes Massif.

The Menderes Massif is generally characterized by Barrovian-type metamorphism of decreasing grade from core to rim, from high-grade and migmatization to greenschist-facies conditions (Bozkurt and Oberhänsli, 2001). This metamorphic feature was first attributed to a single orogenic stage, whose age was initially considered Hercynian based on palaeontologic data (Phillipson, 1918; Önay, 1949; Schuiling, 1962). Later, Ketin (1966) proposed for the first time a late Cretaceous-Paleocene age that he related to the “Laramide orogeny”, while Brinkmann pushed the metamorphism back into the Jurassic (Brinkmann, 1967; 1971). The discovery of upper Cretaceous rudists in emery-bearing marbles of the upper part of the cover sequence (Dürr, 1975) demonstrated the alpine age of the last metamorphism and deformation.

The gneissic core, of late Precambrian age (Hetzel and Reischmann, 1996b; Loos and Reischmann, 1999), was, however, shown to have experienced a polymetamorphic evolution, including two phases of Barrovian-type metamorphism (see for review Candan et al., 2011). The core-cover structure was recently re-emphasized by a detailed study in the Cine sub-massif by Candan et al. (2011) who provide stratigraphic, structural and geochronological evidence for a major unconformity, including basal metaconglomerates, between the core and cover series. They have published detailed maps, where this relationship can be observed. Structures related to late crustal thickening are nappes separated by top-to-the-south thrusts (Ring et al., 1999a; Gessner et al., 2001b). The presence of nappes is sometimes opposed to the core-cover interpretation of the Menderes massif. The presence of Late Cretaceous rudists in marbles of the cover has been used as an argument in favour of the nappes structure. In the southern Menderes Massif the recrystallized limestones with Cretaceous rudists occur on top of the sequence and not below the so-called nappes; see the detailed descriptions in the two papers: Özer, (1998) and Özer et al. (2001). In the central Menderes Massif, they lie below the Permian-Carboniferous sequence due to post-metamorphic thrusting that has disrupted a stratigraphically well-characterized sequence, leaving no place for exotic nappes.

The massif has then been reworked by extensional structures, first shallow-dipping detachments and then steeply-dipping normal faults bounding the present grabens (Bozkurt and Park, 1994; Hetzel et al., 1995a; 1995b; Bozkurt and Park, 1997b; Lips et al., 2001). Recent observations (Bozkurt et al. 2011a; Candan et al. 2011) show that the northern migmatites are related to the exhumation of the Menderes massif below the Simav Detachment and that most of the high-T rocks in the core of the massif are related to the Panafrican orogeny.

Owing to the overlapped metamorphic events in the gneissic core, the maximal P-T conditions reached during the Alpine orogeny remain uncertain. Recent petrologic investigations of Palaeozoic schists placed the peak of the Alpine Barrovian metamorphism (named “Main Menderes Metamorphism”, MMM) into the lower amphibolite facies (Okay, 2001; Whitney and Bozkurt, 2002; Régnier et al., 2003; 2007). Its timing is constrained by, on the one hand, Paleocene foraminifers in the olistostromal formation in the upper part of the cover sequence (Özer et al., 2001) and, on the other hand, lower-middle Miocene sediments at the bottom of the Menderes grabens (Seyitoglu and Scott, 1991 ; Yılmaz et al., 2000). Rb/Sr ages are widely scattered between 63-48 Ma (muscovite) and between 50 and 27 Ma (biotite) (Akkök, 1983; Sengör et al., 1984; Satir and Friedrichsen, 1986; Dora et al., 1990). The MMM has been explained by crustal thickening due to southward transport of the Lycian Nappes and Neotethyan ophiolites away from the plate boundary (Sengör et al., 1984; Okay and Tüysüz, 1999).

This metamorphism was not associated to any significant subduction of the Menderes Massif until the discovery of relict Fe-Mg-carpholite in the cover sequence of the southern massif (Rimmelé et al., 2003b). These H*P*-L*T* metasediments, if they belong to the Menderes Massif, indicate that at least this part of the Menderes Massif was buried to at least 35 km along a cold gradient (up to 12 kbar, 450-500°C) (Rimmelé et al., 2003b; 2005). If, alternatively, the HP metasediments belong to the Cycladic Blueschists, then there is no longer any evidence for a subduction of the Menderes Massif; this point is still under discussion but the lithology of the unit where Fe-Mg-carpholite has been discovered in the cover of the Menderes massif is different from the Cycladic Blueschists and the P-T conditions are also different, with a peak of pressure that is lower. Besides, Fe-Mg-carpholite has never been found in the Cycladic Blueschists probably because of an unsuitable chemical composition (too sodic). The early retrograde stage was dated to ca. 45 Ma (^{40}Ar-^{39}Ar white mica ages) (Pourteau, 2011). Whether the entire Menderes Massif or only its southernmost part experienced this H*P*-L*T* metamorphism remains uncertain, so is its relation with the Barrovian-type stage. However, the preservation of diaspore in metabauxites in this area excludes the possibility that the Barrovian metamorphism significantly affected the southern part of the Menderes massif. One solution is that the HP-LT metamorphism is slightly older than the MMM. It could correspond to the first burial of the Menderes massif in the subduction zone where a cold gradient was preserved before the crust was thickened, which rose the thermal gradient in the Eocene in the deepest parts of the massif.

Fe-Mg-carpholite-bearing sediments were also reported from the northwestern Lycian Nappes near the contact with the Menderes Massif (along the southern and eastern margins as well as in klippen on top of it and the Cycladic Blueschists), while other parts of the Lycian Nappes were not metamorphosed (Oberhänsli et al., 2001; Rimmelé et al., 2003a; 2005). Based on their distribution, these H*P*-L*T* metamorphosed Lycian sediments were distinguished as the Ören Unit (Figure 1A) (Pourteau et al., 2010). Kinematic indicators suggest that the Ören Unit was transported southeastwards (after restoration of the extensional tectonics; see Pourteau, 2011) over the Menderes Massif (Rimmelé et al, 2003a; 2006). The Ören Unit represents the western continuation of the Afyon Zone (Pourteau et al., 2012), a supposedly-greenschist-facies unit (Okay, 1984) actually characterized by the widespread occurrence of Fe-Mg-carpholite (Candan et al., 2005 ; Pourteau et al., 2010). High-pressure metamorphism in the Ören-Afyon Zone took place between 65 Ma (easternmost Afyon Zone) and 60 Ma (Ören Unit) as shown by ^{40}Ar-^{39}Ar white mica ages (Pourteau et al., 2010). To the north and structurally above the Afyon Zone, the Tavşanlı Zone represents the most frontal part of the Anatolide-Tauride microcontinent (Okay, 1984; 1986), which entered first into the subduction zone and was metamorphosed under the H*P*-blueschist-to L*T*-eclogite-facies conditions (Okay and Kelley, 1994 ; Okay, 2002 ; Davis and Whitney, 2006 ; Cetinkaplan et al., 2008) around 88-80 Ma (Okay et al., 1998 ; Sherlock et al., 1999 ; Seaton et al., 2009). Ophiolites sitting atop the Anatolide-Tauride Block generally consists of peridotite (harzburgite with chromite pods), layered and massive gabbros and sheeted dykes (Robertson, 2002). Sub-ophiolitic metamorphic soles, widespread at the region scale yielded clustered radiometric ages of 95-90 Ma (Celik et al., 2006). In the early-middle Eocene, granodiorite inclusions with mantle-derived geochemical signatures intruded the Tavşanlı Zone and overlying peridotite, post-dating ophiolite emplacement in this area (Harris et al., 1994).

2.4 Neogene extension in West-Anatolia

Neogene extension in western Anatolia mainly affected the Menderes Massif, but also the westernmost Pontides in the Kazdağ Massif (Harris et al., 1994; Okay and Satir, 2000). The alpine structure of the Menderes Massif has then been reworked by extensional structures: first shallow-dipping detachments and then steeply-dipping faults bounding the present grabens (Bozkurt and Park, 1994; 1997b; Hetzel et al., 1995a; 1995b; Lips et al., 2001). Among the shallow-dipping extensional structures, the Simav Detachment played a major role in exhuming the northern part of the Menderes massif (Isik and Tekeli, 2001; Isik et al., 2004; Ring and Collins, 2005; Bozkurt et al., 2011b). It separates the Menderes high-temperature gneiss with Oligo-Miocene metamorphic ages below, from the Afyon H*P*-L*T* metamorphic zone. Based on geological markers, Ring et al. (2003) estimated that part of the Menderes Massif exhumation was accommodated in the early Miocene by about 50 km of horizontal extension along the Simav Detachment. Van Hinsbergen (2010), in a recent attempt to reconstructing the progressive exhumation of the Menderes massif also suggested that the early Miocene phase of extension accommodated only 50 km of horizontal extension along the Simav Detachment and that the rest of the Menderes Massif exhumation occurred after 15 Ma by a gliding of the Lycian nappe toward the SE in combination with exhumation of the Central Menderes Massif along the Buyuk Menderes and Alasehir detachments.

2.5 Correlations between the Cyclades and the Menderes Massif

The correlations between the Cyclades and the nearby Menderes massif have been a matter of debate. The main discussion concerns the paleogeographic affinity of the two basements. Basement rocks can be found in the centre of the Cyclades (Naxos, Paros, Ios and Sikinos). According to Ring et al. (1999a) important differences exist between the Menderes and the Cycladic basement. These differences pertain to the age of the basement (Palaeozoic in the Cyclades and Neoproterozoic to Cambrian in the Menderes) and the absence of H*P*-L*T* metamorphism in the Menderes basement (Gessner et al., 2001b; 2004). The second point has been proven wrong by the discovery of H*P*-L*T* parageneses in the Permian cover of the Menderes massif (Rimmelé et al., 2003b) showing that the entire massif or part of it had been involved in the subduction process (see above). For the age of the basement this point is highly debatable. As shown by Keay and Lister (2002) U/Pb zircons ages in the Cycladic basement and metasediments go back to the Palaeoproterozoic with a peak in the Neoproterozoic and the Phanerozoic (in the range 450-400 Ma for the pre-Carboniferous peak). An alternative view was proposed by Jolivet et al. (2004b): the basement and cover of the Menderes massif correlate with the Gavrovo-Tripolitza nappe and its basement. The relations between the two basements may have been complex in the Paleozoic or before, but the presence of an old basement is ascertained in the Cyclades as well and both basements have a similar Mesozoic cover. The important point is thus that the two basements were amalgamated before the deposition of this cover and behaved as a single paleogeographic domain during the building of the Hellenides.

3. The main structures, their ages and their kinematics

Figure 3 shows a detail of the geology of the western domain, centered on the Aegean Sea. The main structures that accommodated shortening during the construction of the Hellenides fold-and-thrust belt (figures 1A and 3) and the internal Rhodope massif were reactivated as exhumation faults and/or extensional detachments.

Finite stretching and crustal thinning appears stronger in the central Aegean than in western Turkey. Moho depth decreases from about 35 km below Central Anatolia to 30 km below the Menderes Massif (Mutlu and Karabulut, 2011) to a maximum of 25 km below the Cyclades and even much less (15 km) below the North Aegean Trough and the Cretan Sea (Makris, 1978; Bohnhoff et al., 2001; Tirel et al., 2004b; Sodoudi et al., 2006). A large part of the finite stretching was accommodated by the North Cycladic Detachment System (NCDS) and the West Cycladic Detachment System (WCDS) (Grasemann et al., 2012) and its possible lateral equivalent in Turkey, the Simav Detachment (see a discussion below). The strong bend between the NCDS and the Simav detachment is compatible with the left-lateral N-S-trending distributed transfer zone between the Aegean Sea and the Menderes Massif proposed earlier by Ring et al. (1999b) and recently re-emphasized by Sözbilir et al. (2011). This zone is also characterized by a surge of alkaline volcanism and even adakites in the Middle Miocene suggestive of a rise of hot asthenospheric mantle there (Pe-Piper and Piper, 2006; 2007; Dilek and Altunkaynak, 2009).

3.1 Active deformation, Plio-Quaternary

A very active N-S extension induces the most active seismicity in Europe (figures 1B; 1G; 2) (e.g. Corinth Rift) (Jackson, 1994; Armijo et al., 1996; Hatzfeld et al., 2000; Lyon-Caen et al., 2004; Bernard et al., 2006). Active extension is recorded today in the prolongation of the North Anatolian fault in the Evia and Corinth Rifts (Armijo et al., 1996). Another roughly orthogonal extension (E-W extension) is observed in the Hellenic arc, southern Peloponnese and Crete, where major normal faults control the topography and produce destructive earthquakes such as the 1986 Kalamata event (Lyon-Caen et al., 1988; Armijo et al., 1992; Jackson, 1994; Goldsworthy et al., 2002). Recent grabens cutting the Menderes Massif also testify for active extension in western Turkey at an overall rate equal to, or higher than, that of western Greece (Eyidogan and Jackson, 1985; Bozkurt and Sözbilir, 2004; Aktug et al., 2009). The SW Aegean and the central Cycladic region, very active in the Miocene, appear seismically inactive and without significant internal strain according to GPS data (McClusky et al., 2000; Floyd et al., 2010; Le Pichon and Kreemer, 2010; Reilinger et al., 2010). The GPS data in the Aegean Sea shows an increase of the stations velocities with respect to Eurasia (figure 1C) compatible with the active extension. Different interpretations of the GPS data are available in the recent literature. All models agree on the rigid motion of the main part of Anatolia about an Eulerian pole north of the Egyptian coast. Then, radically different models seem to equally fit the GPS velocity field in the Aegean Sea: a mosaic of numerous rigid blocks (Nyst and Thatcher, 2004) as opposed to a continuous velocity field in a flowing fluid driven by gravitational forces between the elevated plateau in Eastern Turkey and the oceanic crust in the west (Floyd et al., 2010) or a model using fracture mechanics and a propagating strike-slip fault (Flerit et al., 2004).

The North Anatolian Fault (NAF) appeared quite late in the Aegean framework, not before the Late Miocene, approximately at the time of the Messinian Salinity Crisis (Armijo et al., 1999; Lacassin et al., 2007; Melinte-Dobrinescu et al., 2009). It has accommodated some 80 km of dextral motion since this period during which N-S extension seems to have abandoned the centre of the Cyclades to concentrate in and around the Menderes Massif, as well as between the westernmost tip of the NAF and the Hellenic Trench, within the so-called Central Hellenic Shear Zone (Papanikolaou and Royden, 2007; Royden and Papanikolaou, 2011). The prolongation of the NAF in the Aegean domain is manifested by the North Aegean Trough, a narrow highly subsiding transtensional domain (Lyberis and Deschamps, 1982; Lyberis and Sauvage, 1985; Papanikolaou et al., 2002). How the displacement along the NAF is actually transferred to the west and to the Hellenic Trench is an open question. The total amount of extension within the Central Hellenic Shear Zone is difficult to estimate (Corinth Rift plus Volos Rift plus the normal faults along the northwestern part of the NCDS) as one also has to consider normal faults in the North Aegean Trench, the throw of which is totally unknown. Strike-slip focal mechanisms are also observed south of the North Aegean Trough in a series of splays of the NAF (Figures 1G, 2) (Taymaz et al., 1991; Jackson, 1994; Roumelioti et al., 2011). They disappear southwest of the northeastern coasts of Thessaly and Evia and they are observed again in the vicinity of the Kephalonia Fault, west of the Corinth Rift. It is noticeable that the southwest limit of strike-slip earthquakes in the Aegean Sea coincides with the position of the NCDS, suggesting that this structure may still play a major role in the distribution of deformation at crustal scale, although some conjugate systems of recent faults in Evia also suggest that the propagating NAF has started to affect the crust south of the NCDS (Ring et al., 2007a). However, recent large strike-slip faults related to the NAF are restricted to the thinned crust north of the NCDS and they do not cross the Cyclades archipelago or the extensional domain between Evia and the Corinth Rift (figure 2).

3.2 Pre-Pliocene deformation

Cordilleran-type metamorphic core complexes of Oligo-Miocene age (figures 1A, 1D, 3, 5) are found in the Cyclades archipelago, in the Northern Aegean, the Rhodope Massif in northern Greece and Bulgaria, and the Menderes Massif in Western Turkey (Lister et al., 1984; Avigad and Garfunkel, 1989; Gautier et al., 1993; Sokoutis et al., 1993; Bozkurt and Park, 1994; Gautier and Brun, 1994b; Jolivet et al., 1994a; Hetzel et al., 1995a; 1995b; Ring et al., 1999a; Okay and Satir, 2000; Bozkurt, 2001; Bozkurt and Oberhänsli, 2001; Gessner et al., 2001a; Lips et al., 2001; Isık et al., 2003; Kounov et al., 2004; Bonev et al., 2006; Brun and Sokoutis, 2007). In the Aegean Sea, an Oligo-Miocene H*T*-L*P* evolution (Altherr et al., 1979; 1982; Wijbrans and McDougall, 1986; 1988; Wijbrans et al., 1993; Lips et al., 1998), contemporaneous with extension, overprints a H*P*-L*T* stage coeval with the formation of the Eocene accretionary wedge at the expense of the Apulian continental block and the Pindos ocean (Altherr et al., 1982; Bonneau and Kienast, 1982; Blake et al., 1984; Bonneau, 1984; Lister and Raouzaios, 1996; Vandenberg and Lister, 1996; Foster and Lister, 1999b; 2009). An older (~80 Ma) stage of HP-LT metamorphism was postulated on the basis of dated zircons but recent investigations suggesting that the zircons are magmatic seem to rule out this Cretaceous event (Bulle et al., 2010; Fu et al., 2012). The younger Oligo-Miocene blueschists and eclogites found in the external arc, in Crete and the Peloponnese (Seidel, 1978; Seidel et al., 1982; Theye and Seidel, 1991; 1993; Theye et al., 1992; Jolivet et al., 1996; 2010c) formed and were exhumed within the same time frame as the H*T*-L*P* core complexes further north during backarc extension (Jolivet et al., 1994b; 2004a). All metamorphic units, whether H*P*-L*T* or H*T*-L*P*, were exhumed below crustal-scale detachments faults and extensional ductile shear zones, in the backarc domain as well as in the accretionary wedge (Jolivet et al., 2003; Ring and Layer, 2003; Brun and Faccenna, 2008; Ring et al., 2010; 2007b). Still older H*P*-L*T* metamorphic rocks are found also in the Rhodope and Peri-Rhodope areas, the peak of pressure dating back to the Jurassic and Cretaceous (Kostopoulos et al., 2000; Mposkos and Kostopoulos, 2001; Burg, 2011; Nagel et al., 2011). The domain of the Oligo-Miocene extension was bounded in the northeast by the dextral Eskisehir fault with a total offset of 100 km (Okay et al., 2008). During the Oligocene, the Eskisehir Fault had a similar function as the present North Anatolian Fault by moving crustal material toward the Aegean domain.

3.3 Age of backarc extension in the Aegean region

The age of the first backarc extension is still a disputed point. In early interpretations, it was supposed to start as late as 13 Ma (Le Pichon and Angelier, 1981b) or even 5 Ma (McKenzie, 1978). Field observations however suggested earlier ages in the Early or Middle Miocene (Mercier et al., 1976). The earliest unconformable marine deposits in supra-detachement basins date from the base of the Miocene (Aquitanian 23-20 Ma), they are found in Naxos and Evia (Guernet, 1971; Angelier et al., 1978; Katsikatsos et al., 1981; Sanchez-Gomez et al., 2002; Kuhlemann et al., 2004). The base of the Miocene could then be considered as the earliest possible date for extension (Ring et al., 2010). However, the base of the Miocene is the time when the topographic surface of the crust of the Hellenides locally reached sea level during thinning, and extension must have started earlier. Another type of argument comes from P-T-t paths. The high-temperature domes of the central Cyclades (Naxos-Paros and Mykonos) are commonly associated with extension and crustal thinning (Urai et al., 1990; Buick, 1991; Gautier et al., 1993; Vanderhaeghe, 2004). The core of the Naxos dome is made of migmatites where zircons partly preserve the Paleozoic ages of the original Cycladic basement and show younger rims, suggesting that partial melting occurred in the Early Miocene prior to 20.7 Ma (Keay et al., 2001). Thermobarometry and Rb/Sr dating further suggest that the dome formed between 20 and 15 Ma (Duchêne et al., 2006). This in fact gives a minimum age for the initiation of extension in the area, as partial melting and the formation of the dome can occur only after some thermal relaxation of the orogenic crust has happened and nothing says how long have the conditions for partial melting been present. Exhumation paths in lower temperature metamorphic rocks may give a more precise answer as they better preserve the details of the exhumation history. In the case of Mount Olympos $^{40}Ar/^{39}Ar$ ages on phengites suggest a cooling below 100-150°C at 16-23 Ma, a date that is interpreted as the inception of the present-day system of steep brittle normal faults that bounds the massif from the Thermaikos Gulf to the east (Schermer et al., 1990; Lacassin et al., 2007). This, again, gives a minimum age for the beginning of extension. The example of Tinos island shows a two-staged exhumation path (Parra et al., 2002): from 16-18 kbar to ~9 kbar the Cycladic Blueschists were exhumed along a HP-LT path between ~45 Ma and 37 Ma. Between 37 and 30 Ma isobaric heating is recorded, followed by renewed exhumation along a warmer path until the middle Miocene. The first part of the exhumation path was attributed (Parra et al., 2002) to syn-orogenic exhumation and the second path to post-orogenic extension. The renewed exhumation would be a consequence of the inception of backarc extension. The isobaric heating corresponds to a period when the CBS were no longer in the subduction channel and not yet subjected to backarc extension. This interpretation sends the inception of extension back to the Early Oligocene at least. Further north, in the Rhodope massif, extension has started even earlier. The first ascertained extension is recorded in the Mesta basin where late Eocene (Bartonian, 40-37 Ma) deposits are clearly related to normal faulting (Burchfiel et al., 2003; Georgiev et al., 2010; Burg, 2011). Extension reached a maximum in the Bartonian, contemporaneouly with a surge of calc-alkaline volcanism. Bartonian is also the age of the first formation of the Thrace Basin east of the Rhodope Massif (Okay et al., 2010). The main detachments of the Rhodope region such as the Tokachka detachment in the Kesebir-Kardamos dome (within the Rhodope) (Bonev et al., 2006) stopped their activity some 33 Ma ago before they were cut by high-angle normal faults (Wüthrich, 2009; Burg, 2011). The syn-tectonic Kavala and Vrondou plutons show a C/S fabric related to the normal shearing along the Strymon and Kavala detachments; they are dated respectively at 21 and 30 Ma attesting an Oligocene and Early Miocene age of extension (Kolocotroni and Dixon, 1991; Dinter and Royden, 1993) contemporaneous with the second exhumation of the Tinos core complex below the NCDS. In the same region, the gneiss of Thasos island were exhumed below a detachment between 26 and 8 Ma (Sokoutis et al., 1993; Wawrzenitz and Krohe, 1998; Brun and Sokoutis, 2007). There is thus ample evidence that extension started much earlier than the Early Miocene in the northern part of the Aegean region. When considering the age and location of magmatic products in the whole Aegean region a drastic change is observed at ~35 Ma ago. Before 35 Ma the magmatic centres were located in the Balkans and part of the Rhodope and they moved southward after 35 Ma at a constant rate of ~3 cm/yr (Jolivet et al., 1998; Jolivet and Brun, 2010). All these arguments suggest a two-staged scenario with a first period of extension in the Rhodope before 35-33 Ma without migration of the volcanic centres and a second stage after this period, with a fast migration that can be interpreted as a consequence of slab retreat. The first stage of extension is interpreted as a consequence of lithospheric delamination and heating of the crust by the uplifted asthenosphere (Burg, 2011). Backarc extension has thus begun some 30-35 Ma ago in the northern Aegean region as in other Mediterranean backarc basins (Jolivet and Faccenna, 2000).

3.4 Age of thrusting in the northern Cyclades and Evia island

This question of the age of the first extension is linked to another important question on the age of the last thrusting and related H*P*-L*T* metamorphism in the northern Cyclades and further to the NW until Mount Olympos. The age of thrusts is constrained by (1) the age of the youngest sediment in the footwall unit, that gives a maximum age, and the radiometric ages of syn-kinematic minerals. From Mount Olympos to Evia several tectonic windows (Olympos, Ossa, Almyropotamos) show a metaflysch resting on top of the footwall unit (Gavrovo-Tripolitza nappe) overthrust by the Cycladic Blueschists or their equivalent (Ambelakia unit) or the Pelagonian (Godfriaux, 1962; Guernet, 1971; Dubois and Bignot, 1979; Ferrière, 1982; Godfriaux and Ricou, 1991). This metaflysch contains nummulithes that constrain the deposition ages from the Lower to Middle Eocene (Ypresian-Lutetian in Dubois and Bignot, 1978), at the youngest 40 Ma (and not Eocene-Oligocene as mentioned in Ring et al. (2007a)). Usually, flysch are deposited immediately before their burial below the orogen. This pleads for an Eocene age of thrusting. Radiometric ages (^{40}Ar-^{39}Ar) in the Cycladic Blueschists or Ambelakia unit are indeed Eocene (Schermer et al., 1990) and the reversal of shear sense started some 33 Ma ago in Evia (Ring et al., 2007a). It can be debated whether this reversal of shear sense corresponds to the beginning of exhumation as an extruding wedge (Ring et al., 2007a) or to the inception of extension. As a matter of fact, 33 Ma is approximately the age of the beginning of the Aegean extension as discussed above. So we prefer a solution where thrusting is essentially Eocene.

3.5 Exhumation below post-orogenic detachments and syn-orogenic exhumation shear zones

The terminology of detachments and exhumation shear zones is sometimes ambiguous and we wish to reemphasize here that the presence of a detachment does not necessarily means whole crust extension. Some detachments result from extension at the scale of the lithosphere driven by extensional boundary conditions and they lead to crustal thinning. In this category the most emblematic examples are the Basin-and-Range detachments (Wernicke, 1992) or the Aegean one, such as the NCDS. Alternatively, some detachments accommodate the exhumation of deep metamorphic rocks at the top of a growing wedge or a subduction channel and they do not lead to crustal thinning (Jolivet et al., 2003; Ring and Glodny, 2010; Ring et al., 2010). The most spectacular example is the South Tibetan Detachment (Burg et al., 1984) at the top of the extruding high-temperature gneiss of the High Himalayas. We then refer to syn-orogenic detachments (Jolivet et al., 2003) (or exhumation faults) as opposed to post-orogenic detachments in the case of whole crust extension.

The NCDS (Mehl et al., 2005; 2007; Jolivet et al., 2010b; Lecomte et al., 2010; Denèle et al., 2011) has exhumed the Cycladic metamorphic rocks from the Eocene when it acted as an exhumation fault at the top of the Eocene subduction channel, up to the Miocene when it was reactivated as a Basin-and-Range-type detachment in a whole-crust extensional setting. Syntectonic plutons were emplaced during the activity of those detachments (Faure and Bonneau, 1988; Faure et al., 1991; Lee and Lister, 1992; Lister and Baldwin, 1993; Kumerics et al., 2005; Brichau et al., 2006; Brichau et al., 2007; 2008; 2010; Bolhar et al., 2010; Jolivet et al., 2010b; Denèle et al., 2011). A series of southwest-dipping low-angle normal faults (LANF) were recently described in the southwest Cyclades with shearing directions similar to those of the northern and central Cyclades, but with an opposite shear sense. This set of LANF is described as the Western Cycladic Detachment (WCDS) (Iglseder et al., 2011; Grasemann et al., 2012). Ring et al. (2011) have defined recently the South Cycladic Detachment System (SCDS) on Sifnos, a probable extension of the WCDS. A top-to-the-South detachment was described in Ios in the southern Cyclades (Lister et al., 1984; Foster and Lister, 1999a) but a recent reexamination of field relations suggests that this structure is a syn-H*P* thrust of Eocene age preserved at the base of the Cycladic Blueschists reactivated by top-to-the-north extensional shear afterward (Huet et al., 2009; Huet, 2010). The age of the beginning of top-to-the-south shearing has been estimated to ~35 Ma (this is in fact a minimum age) and the end at ~30 Ma (Foster and Lister, 2009). Top-to-the-north ductile shear stopped at ~18 Ma and the exhumation in the brittle domain continued until ~9 Ma (Thomson et al., 2009).

Further east, the Simav Detachment (SDF) (Isik and Tekeli, 2001; Isık et al., 2003; 2004; Ring et al., 2003; Ring and Collins, 2005; Thomson and Ring, 2006; Ersoy et al., 2010b; Bozkurt et al., 2011a; 2011b) has exhumed the northern part of the Menderes Massif, where Oligo-Miocene H*T*-L*P* metamorphism and migmatites are observed. The hangingwall of the SDF belongs to the Afyon Zone (Pourteau et al., 2010). Radiometric ages on the metamorphic and intrusive rocks (muscovite and biotite Rb-Sr, zircon and apatite fission track and (U-Th)/He ages) indicate that this fault was active from 30 to 12-8 Ma (Ring et al., 2003; Ring and Collins, 2005; Thomson and Ring, 2005; Ersoy et al., 2010b; Bozkurt et al., 2011), that is exactly the same period as the NCDS. Besides, the Simav Detachment has exhumed Oligo-Miocene high-temperature metamorphic cores just south of the suture zone, exactly like the NCDS. The exact amount of displacement along the Simav detachement is debatable but several tens of kilometers is a likely figure. Based on geological offset markers it has been estimated to 50 km (Ring et al., 2003). The same value has been used for reconstructions of the Menderes massif by van Hinsbergen (2010) and Pourteau et al. (2010) while 50-70 km is estimated for the NCDS (Jolivet et al., 2004a; 2010b). Its general setting, its location with respect to the Vardar suture, the finite amount of displacement, the timing of its activity and the fact that it corresponds to a reactivated crustal-scale thrust thus suggest that the Simav Detachment may be an extension of same crustal-scale structure as the NCDS.

Rb/Sr muscovite, biotite and apatite fission-track ages indicate that the Selale detachment in the Kazdag massif was active in the Early-Middle Miocene (21 to 14 Ma) (Okay and Satir, 2000; Beccaletto and Steiner, 2005; Cavazza et al., 2009) and led to the exhumation of the amphibolite facies gneisses and amphibolites. Younger detachments formed during the Miocene within the Menderes Massif and they were recently replaced by steeper normal faults that shaped the Menderes E-W trending grabens (Hetzel et al., 1995a; Gessner et al., 2001a; Lips et al., 2001; Bozkurt and Sözbilir, 2004). A preserved top-to-the-east syn-orogenic detachment or exhumation shear zone, active during the Eocene, is found in the Cyclades on Syros, where the CBS are topped by the Vari unit constituted of amphibolites and a Late Cretaceous granitic and gneissic basement (Maluski et al., 1987; Trotet et al., 2001a; 2001b; Huet, 2010).

During an earlier period, the Ören unit that preserves H*P*-L*T* parageneses of Danian age (62-59 Ma) (Ring and Layer, 2003; Pourteau, 2012) where exhumed along a top-to-the-southeast shear zone (after reconstruction, Pourteau, 2011) affecting also the underlying marble cover of the Menderes Massif, suggesting the presence of a thick exhumation shear zone (Rimmelé et al., 2003a), between the Ören Unit and the Menderes Massif. The geometrical relations between this exhumation shear zone and a series of top-to-the-south shear zones (South Menderes Shear Zone), deeper in the structure, between the schist cover and the gneissic core of the Menderes Massif (Bozkurt and Park, 1994; 1997a; Régnier et al., 2003) are unclear. Most indications suggest that these are of Eocene age (Hetzel and Reischmann, 1996a), and thus younger than the HP-LT metamorphic event and the exhumation of the Ören unit, and that they are thrusts rather than detachments. An Eocene thrusting event would be partly contemporaneous with renewed thrusting in the Lycian Nappes further south (Collins and Robertson, 1998). The Ören Unit would then be transported by these thrusts after most of its exhumation. In this interpretation the southern limit of the Menderes Massif would not be a Cenozoic detachment. Only the northern part of the massif would have been exhumed below a detachment (the Simav Detachment) and the southern part would have been exhumed earlier (Gessner et al. 2001a; see also (van Hinsbergen, 2010)).

A partly preserved younger (Oligo-Miocene) syn-orogenic detachment (the Cretan Detachment) is observed at the top of the Phyllites-Quartzites (PQ) nappe in Crete and the Peloponnese (Jolivet et al., 1996; 2003; 2010c; van Hinsbergen and Meulenkamp, 2006; Jolivet and Brun, 2010; Zachariasse et al., 2011). It has been reworked in the Peloponnese by younger detachments and then by the active normal faults that shape the relief of the Parnon and Taygetos ranges (Armijo et al., 1992; Papanikolaou and Royden, 2007; Papanikolaou et al., 2009). A still younger detachment or extensional decollement (the Zaroukla decollement) is observed south of the Corinth Rift (Jolivet et al., 2010a) that was cut some 0.6 Ma ago by the recent and active normal faults that control the geometry of the rift (Rohais et al., 2007a; 2007b).

The exhumation of the Southern Rhodope core complex (SRCC) has been first dated to late Oligocene-lower Miocene and attributed to a SW-dipping detachment (Dinter and Royden, 1993; Sokoutis et al., 1993; Wawrzenitz and Krohe, 1998; Bonev et al., 2006; Jahn−Awe et al., 2010). However, evidence for middle Eocene extension in southern Bulgaria (Burchfiel et al., 2003; Burg, 2011) as well as geochronological evidence for middle to late

Eocene cooling in the southern and northern borders of the SRCC indicate that exhumation started in middle Eocene and that it was accommodated by the Kerdylion detachment (Brun and Sokoutis, 2007). Fission-tracks dating (Wüthrich, 2009) confirm this interpretation. In the southern Rhodope core complex, both thrusting (up to Paleocene) and extension (from middle Eocene to lower Miocene) display the same top-to-the-SW sense of shear. Consequently, at some particular places, it is rather difficult to make a clear distinction between structures that are related to either thrusting or extension. This is exemplified by the Nestos shear zone at the northern limit of the SRCC that is interpreted either as the base of a collapsing wedge (Nagel et al., 2011) or the complex product of syn-collision exhumation during Cretaceous-Eocene later reworked by backarc extension (Burg, 2011). A second stage of exhumation was controlled by two sets of steeply dipping normal faults trending NW-SE and NE-SW associated with the deposition of sedimentary basins (Brun and Sokoutis, 2007). This late tectonic extensional event, which was superposed to the core complex history, started in middle Miocene as indicated by fission-tracks ages between 17 Ma and 8 Ma in the vicinity of basin bounding faults (Wüthrich, 2009) and by the Serravalian age of the oldest sediments (Kousparis, 1979).

3.6 Kinematic indicators

Figure 5 shows a synthesis of available kinematic indicators on the main shear zones. We have distinguished them by their age (Eocene or pre-Eocene versus Oligo-Miocene) and tectonic context (thrust –blue-, exhumation shear zones –green-, extension –red-).

Only a few regions undoubtedly show thrust-related stretching lineations. They are all of pre-Eocene and Eocene age. The basal contact of the Cycladic Blueschists in Ios is a syn-H*P* thrust with a top-to-the-south shear sense (Huet et al., 2009). Prograde top-to-the-south kinematic indicators are also found within the H*P*-L*T* nappe pile on Syros island (Philippon et al., 2011). They were preserved by progressive localisation of deformation during exhumation along the contacts between units (Trotet et al., 2001a; Keiter et al., 2004). Syn-H*P* top-the-southwest kinematic indicators can be observed within the Ambelakia unit (equivalent to the CBS) above the Mount Olympos marbles (Schermer, 1990; 1993; Schermer et al., 1990; Godfriaux and Ricou, 1991). Further east, syn-H*P* kinematic indicators on Samos island are related to thrusting of the CBS onto an equivalent of the Gavrovo-Tripolitza nappe (Ring et al., 1999b) and dated Eocene (Ring and Layer, 2003).

Rosenbaum et al. (2007) have proposed the existence of a detachment on Amorgos island, fission-tracks data showing that the exhumation of the lower unit is Early Miocene (Ring et al., 2009), but a more recent study has instead shown thrusting deformation at the contact between the Cycladic Blueschists and an equivalent of the cover of the Menderes Massif, with an NW-SE direction and a possible top-to-NW vergence of thrusts (Chatzaras et al., 2011). The age of this deformation is debatable. The nappe emplacement has to be younger than the Upper Eocene flysch and some brittle deformation was still active in the Early Miocene. All regions show H*P*-L*T* conditions during this top-the-south or –southwest shearing event, at variance with the Eocene top-to-the-south shearing deformation along the contact between the Menderes core series and cover series that is associated with H*T*-L*P* metamorphic conditions (Bozkurt and Park, 1997a).

As discussed by Brun and Sokoutis (2007) the lineation observed in the Rhodope massif is composite and relates partly to the early thrusting episode and partly to the late extensional one that is predominant and has reoriented earlier structures. There, the shearing direction during thrusting is consistent with that observed in the Mount Olympos region.

Kinematic indicators related to Eocene normal shear zones seem more E-W than those related to thrusting. E-W directions and top-to-the-East shear senses are observed on Syros and Sifnos islands, clearly related to the exhumation segment of the *P*-*T* paths and dated from the Eocene (Trotet et al., 2001a; 2001b). Just north of these two islands syn-H*P* NE-trending stretching lineations are preserved locally below the NCDS on Tinos island (Gautier and Brun, 1994b; 1994a; Jolivet and Patriat, 1999), but the late extensional episode has been intense and the observed directions have likely been reoriented. Locally, however, ENE-WSW directions are preserved at a certain distance from the NCDS (Aubourg et al., 2000). In contrast, syn-blueschist facies lineations on the small island of Iraklia south of Naxos, possibly related to exhumation trends N-S (Behrmann and Seckel, 2007) showing that the Eocene pattern is still partly unclear.

The good preservation of H*P*-L*T* parageneses within the Ören Unit implies a fast exhumation soon after the peak of pressure (Rimmelé et al., 2003a). This would imply a Late Cretaceous or Early Paleogene age for the beginning of exhumation. However Ring et al (2007b) have obtained Eocene ages (42-32 Ma) from the Selçuk nappe and the Selçuk extensional shear zone with top-to-the-northeast kinematic indicators. The Ören unit metapelites and marbles are lithologically quite different from the ophiolitic melange of the

Selçuk nappe that resembles more typical Cycladic Blueschists units. Ring et al. (2007b) interpreted this Eocene deformation as an extensional shearing at the top of an extruding wedge, contemporaneous of thrusting at the base of the wedge. This is in good agreement with the interpretation of the exhumation of the Cycladic Blueschists between a basal thrust (Ios) and a detachment (Vari) at the top (Jolivet et al., 2003; Huet et al., 2009; Jolivet and Brun, 2010).

The Oligocene-Miocene period shows a simpler pattern. Syn-exhumation kinematic indicators, related to the Cretan Detachment, are found within the PQ nappe in Crete and the Peloponnese (Jolivet et al., 1996; 2010c; Doutsos et al., 2000; Xypolias and Doutsos, 2000; Papanikolaou and Vassilakis, 2010). Once the Miocene clockwise rotation of the Peloponnese (Kissel and Laj, 1988; Kissel et al., 2002; van Hinsbergen et al., 2005b) has been substracted, all directions are N-S and the shear senses are top-to-the-north in Crete and bivergent in the Peloponnese. From the Cyclades to the Menderes and the Rhodope, all post-orogenic directions are N-S or NE-SW both in the ductile and brittle fields (Mehl et al., 2005). Some of these stretching lineations have been later rotated. Paleomagnetic data (Morris and Anderson, 1996) show that, once the late rotation has been restored, the shearing direction is almost everywhere NE. The case of Mykonos has been reconsidered recently and the N60°E average shearing direction correspond to the clockwise rotation of an initially N30°E stretching lineation during exhumation, preserved only in the vicinity of the NCDS (Denèle et al., 2011). Kinematic indicators are everywhere top-to-the-SW in the Rhodope, bivergent in the Menderes Massif with a NE-SW direction after corrections of paleomagnetic rotations (Van Hinsbergen, 2010), top-to-the-southwest in the western Cyclades and top-to-the-north or –northeast in the northern and central Cyclades. The transition between the top-south and top-north domains is unclear. Our recent observations in the islands of Folegandros and Sikinos reveal consistent top-north kinematic indicators in the CBS, yet undated, but with a clear link with the pervasive retrogression of the H*P* parageneses in the greenschists facies, suggesting an Oligo-Miocene age by comparison with the nearby island of Ios (Huet et al., 2009). The relations between the NCDS and other top-to-the-north detachments and the contemporaneous WCDS are not clear. Solving this question is however important to better constrain the kinematics of domes and detachments and the pattern of strain at the scale of the crust (Grasemann et al., 2012).

4. Mantle structures

Tomographic studies have unravelled the geometry of the Hellenic slab (Spakman et al., 1988; Wortel and Spakman, 1992; 2000; Spakman and Wortel, 2004; Van Hinsbergen et al., 2010). Large-scale studies show a single 1500 km long slab penetrating the lower mantle and a major lateral tear below western Turkey, described as a STEP-fault (Subduction-Transform Edge Propagator) by Govers and Wortel (2005) as suggested by tomographic models (de Boorder et al., 1998; van Hinsbergen et al., 2010). The same authors suggest the presence of an additional STEP-fault below the Kephalonia fault and Corinth Rift in the transitional domain between oceanic and continental subduction. We show images from another P-wave tomographic model (Faccenna et al., 2003; Piromallo and Morelli, 2003) with a resolution comparable to that of the Utrecht team (Spakman et al., 1988), illustrating basically the same geometry (figures 1E, 6). Results from teleseismic scattered waves along a 2-D transect (Suckale et al., 2009) are consistent with the intuition of Govers and Wortel (2005) of a tear below the Kephalonia Fault. This geometry is further confirmed by Gesret et al (2011) who analysed teleseismic converted waves and showed that the top of the slab is offset horizontally by some 100 km on either sides of the supposed tear. Two recent studies provide more detailed images of the slab geometry below Turkey and the Aegean Sea, Biryol et al. (2011) based on P-waves for the whole upper mantle and Salaün et al. (2012) for the 80-300 km depth range based on surface waves: the basic features are preserved but the torn slab below Western Turkey appears to be connected at depth with the Hellenic slab below northern Greece, thus giving an image partly different from a classical STEP-Fault. The two tears, east and west of the Aegean domain thus seem to be confirmed by recent tomographic studies.

Seismic anisotropy investigations show a very simple pattern for SKS waves that sample mainly the fabric of the asthenosphere (figure 1F). Seismic anisotropy compared to crustal deformation (figure 7) is now widely used to discuss crust/mantle coupling/decoupling processes (Barruol and Granet, 2002; Little et al., 2002; Lucente et al., 2006; Buontempo et al., 2008; Jolivet et al., 2009). Splitting measurements of teleseismic shear SKS shear waves give the azimuth of the fast wave polarization and the delay time between fast and slow waves. The direction of fast polarization is considered a proxy for the orientation of mantle deformation. The most significant contribution to anisotropy for SKS waves is due to the upper mantle (mostly the asthenosphere) and only ~10% to the crust (Silver, 1996; Savage, 1999). Comparison between the fast polarization direction of split SKS shear waves splittings and stretching directions measured in exhumed middle-lower crust metamorphic core complexes in the Mediterranean backarc domains suggests that anisotropy data, directly related to the crystallographic fabric, indeed show the horizontal projection of the long axis of the strain ellipsoid in the mantle, that may correspond to flow direction in case of a simple flow pattern (Jolivet et al., 2009).

After the pioneer work of Hatzfeld et al. (2001) that showed a NE-SW fast direction in the mantle below the northern Aegean Sea, recent works (Paul et al., 2010 ; Evangelidis et al., 2011) show a very simple pattern (Figures 1F and 7) that suggests asthenospheric flow toward the Hellenic trench below the backarc region, including the Aegean Sea and the Anatolian plate. Fast directions trend NE-SW below most of Anatolia and they rotate progressively in Anatolia and abruptly in the Aegean Sea, to become more NW-SE below the Hellenides with a delay decrease from the Western Aegean Sea to the Hellenides. Fast directions strike parallel to the subduction zone in the vicinity of the Hellenic Trench and below the continental Hellenides. It is noticeable that the NAF does not deflect the direction of the SKS anisotropy and thus of the asthenospheric flow. Kreemer et al. (2004) and Jolivet et al. (2009) have shown that the direction of mantle flow given by SKS wave splitting is parallel to the direction of crustal stretching shown by field studies, not only in the Aegean MCC but also in the Tyrrhenian and Alboran ones, suggesting that crustal deformation in the backarc region is coupled to the underlying mantle flow through slab and trench retreat. Moreover, the variation of delay times can be correlated with the intensity of finite stretching as well as with large-scale rotations in the upper crust like in the western Aegean region (Brun and Sokoutis, 2010). According to Endrun et al. (2011) Rayleigh waves azimuthal anisotropy shows a more complex fabric within the Aegean lithosphere than in the asthenosphere (figure 7): the lithospheric mantle displays a strong anisotropy in the Northern Aegean with a N-S fast direction and a much weaker anisotropy below the Cyclades and the volcanic arc, while the lower continental crust shows a NE-SW fast direction roughly parallel to the stretching directions of the MCCs south of the North Aegean Trough and a N-S direction north of it.

Pn-waves sample the uppermost lithospheric mantle, just below the Moho discontinuity. The seismic anisotropy they reveal (Mutlu and Karabulut, 2011) shows a trend similar to that of SKS waves below the Aegean Sea, Southwestern Turkey or the Hellenic chain. Below the Cyclades and the Cretan Sea the trend is similar for the uppermost mantle and the lower crust, but different in the lithospheric mantle (Rayleigh waves) suggesting that the lower crust and uppermost mantle are partly coupled, and decoupled from the rest of the lithospheric mantle.

5. Geodynamic evolution

5.1 Insights from the magmatism

The chemical evolution of arc magmatism is strongly controlled by the southward migration of the subduction zone and retreat of the slab as shown by the age of volcanic products (Fytikas et al., 1984; Jolivet and Brun, 2010). After the formation of I-type plutons from Eocene to middle Miocene age that originate from mantle-derived magmas with a strong influence of crustal melts and variable sources, Pe-Piper and Piper (2006; 2007) distinguish five major groups of volcanic products (Figure 8). (1) Arc calc-alkaline volcanism with minor tholeiitic component characterise the Aegean arc and some volcanic rocks in Thrace during the Plio-Quaternary. In the recent volcanic arc, greater contributions of the lithospheric mantle and the crust are observed in the western part than in the central and eastern part. (2) Shoshonitic volcanic products seem to derive from a subcontinental lithosphere enriched by past subducted material. They are found in the northern Aegean and western Anatolia in the Early and Middle Miocene. The most likely interpretation relates them to lithospheric mantle delamination. (3) Backarc basalt, rhyolite and trachyte volcanc rocks, showing many characteristics of shoshonites, are present in small volumes and their source is enriched in sub-continental lithosphere, and they are spatially and genetically associated with subalkaline rocks. Backarc basalts form essentially in the eastern Aegean and western Anatolia from the Middle Miocene to the Quaternary while trachytic rocks also form in the western Aegean and western Rhodope from the Late Miocene. (4) Some high-Mg rocks in Evia, Skyros and Chios show affinities with adakites. The geographic position of shoshonites, adakites and alkali-basalts also seem to fit a delamination event in the Middle Miocene. Backarc basalts are interpreted as the result of local thermally induced mantle melting during crustal thinning in the backarc region of a delaminated slab. (5) Alkali volcanism appears in Western Anatolia in the Plio-Quaternary.

Dilek and Altunkaynak (2009) and Ersoy et al. (2010a) describe the magmatic evolution of western Turkey as follows. The first magmatic episode corresponds to Eocene granitoids and related volcanic products north and south of the Izmir-Ankara suture zone. This arc extends westward in the Rhodope and the Balkan. It corresponds to the late evolution of the Late Cretaceous subduction volcanic arc that developed above the subduction of the Vardar ocean (Marchev et al., 2005; von Quadt et al., 2005). The second episode shows a wide distribution of Oligo-Miocene volcanic and plutonic rocks with medium to high-K calc-alkaline compositions, associated with an increasing crustal component through time. It is interpreted as a consequence of post-collision slab break-off that brought the heat to melt the mantle lithosphere below the suture zone, or as an episode of lithospheric delamination (Burg, 2011). The next episode is made of alkaline bimodal volcanism in the Middle Miocene with an important contribution of asthenospheric mantle-derived melts during partial delamination of the lithospheric mantle during roll-back. From ~12 Ma started the last episode with uncontaminated highly alkaline volcanic rocks derived from an asthenospheric mantle melt. The southward migration of the N-S trending volcanism of western Turkey from 21 to 4 Ma is further interpreted as a consequence of the progressive formation of the STEP-Fault postulated on the basis of tomographic models by Govers and Wortel (2005) and de Boorder et al. (1998).

In both descriptions the importance of lithospheric mantle delamination and the influence of asthenospheric melts are emphasized. With the addition of the eastern Turkey alkalic to peralkalic volcanism related to lithospheric delamination below the collision zone (Gögüs and Pysklywec, 2008a; 2008b; Sengör et al., 2008; Gögüs et al., 2011) it shows that this process was common in the mantle below the Anatolia-Aegea region throughout the Cenozoic.

5.2 Insights from metamorphism

Metamorphic soles of obducted ophiolite in the Hellenides and the Anatolides-Taurides display different ages. While in Greece, south of the Vardar suture, all ages are Late Jurassic, most of them are late Cretaceous (95-90 Ma) in Turkey atop the Anatolides-Taurides (see a review by Parlak and Delaloye, 1999) with only one exception northeast of Ankara that shows a late Jurassic age interpreted as an intra-oceanic subduction (Celik et al., 2011). The obduction event is indeed Late Jurassic to Early Cretaceous in the Hellenides and further north in the Dinarides (Schmid et al., 2008) and Late Cretaceous in Turkey (Okay et al., 2001), but also in the Vardar suture zone. Late Cretaceous amphibolites can be found also in the Cyclades, as for instance in Tinos island (Maluski et al., 1987; Katzir et al., 1996) suggesting that the late Cretaceous suture zone is continuous below the Aegean Sea from the Vardar to the Izmir-Ankara suture.

Late Jurassic-Early Cretaceous metamorphic ages (170-120 Ma) were obtained from UH*P* metamorphic rocks in the Rhodope massif and their geodynamic significance is not ascertained (Burg, 2011). This metamorphic event is contemporaneous with a period (155-130 Ma) of contraction, metamorphism and cooling in the Strandja orogenic belt, to the NE of the Rhodope (Sunal et al., 2011). High-pressure conditions are recorded in the Rhodope in the early Cretaceous (130-115 Ma) before a first extensional episode contemporaneous with amphibolite-facies and greenschist metamorphism from 110 to 65 Ma. The Rhodope orogen was then eroded and the first sediments deposited some 60 Ma ago. Significant orogenic events are then delayed until the Eocene when the Cycladic Blueschists were formed further to the south (see a synthesis in Burg (2011)).

Less disturbed by late extension, the Anatolide-Tauride belt reveals a clearer distribution of sutures zones and HP-LT metamorphic belts than the Cyclades. Late Cretaceous blueschists and eclogites are also found further north in the Biga peninsula and the Dardanelles Strait (Okay et al., 2001; Topuz et al., 2008). In Greece, Late Cretaceous ages are also found in H*P*-L*T* metamorphics from the Pelagonian domain (Lips et al., 1998). There is thus a large late Cretaceous belt of HP-LT metamorphic rocks in the vicinity of the Vardar – Izmir-Ankara suture zone that pleads for continuous accretion south of the Rhodope during this period.

Aegean H*P*-L*T* metamorphic units, the Eocene Cycladic Blueschists or the Oligocene and Early Miocene Phyllite-Quartzite Nappe, were exhumed during slab retreat. Most of the deformation seen today in the field within these units is related to exhumation and not to burial. The major part of exhumation was achieved within the contact between the two plates as an extrusion wedge within the subduction channel (Jolivet et al., 2003; Brun and Faccenna, 2008; Huet et al., 2009; Jolivet and Brun, 2010; Ring et al., 2010). Exhumation was accommodated by contemporaneous thrust shear zones at the bottom of the wedge and extensional shear zones and normal faults at the top of the wedge. The latter contacts are preserved today as syn-orogenic detachments below the Upper Cycladic Unit on Syros or the Gavrovo-Tripolitza nappe in Crete and the Peloponnese. Very little is usually left of the prograde deformation history because exhumation involves intense shearing. Some prograde top-South kinematic indicators have been recently described within the eclogite-blueschist Kastri unit on Syros island (Philippon et al., 2011) but in most cases only retrograde features can be observed and analysed on a scale useful to draw conclusions in terms of tectonic processes (Trotet et al., 2001a). *P-T*-time paths also show most of the time only the retrograde episode (Avigad and Garfunkel, 1991; Avigad et al., 1992; 1997; Avigad, 1998; Trotet et al., 2001a). The analysis of such *P-T*-time paths in the Cyclades and the Phyllite-Quartzite Nappe shows that the syn-orogenic exhumation (before the whole crust thinning) is achieved along cold paths, implying an evolution within the subduction channel. The case of the Phyllite-Quartzite Nappe shows an evolution along strike from colder retrograde paths in the east (east and central Crete) toward warmer paths in the west (Pelopponese) (Trotet et al., 2006; Jolivet et al., 2010c). This evolution can be associated with an evolution of kinematic boundary conditions from the Cretan transect where the channel is opened by slab retreat and *P*-*T* paths are colder to the northern Peloponnese one where the channel is tighter because of a less efficient retreat (see also (Beaumont et al., 1999).

6. Tectonic synthesis, progressive localisation of deformation

6.1 The Rhodope Massif, the internal Hellenides and the Cyclades

North of the Vardar suture zone, a protracted period of accretion, associated with H*P*-L*T* or UH*P*-L*T* metamorphism, is recorded from the Early Cretaceous to the Late Cretaceous. In the Rhodope Massif, the major Nestos shear zone that juxtaposes the Rhodope terrane of mixed continental and oceanic origin with HP and UHP overprint on top of the Pangaion-Pirin complex, a Variscan basement and its cover, has been later cut by the Kerdylion Detachment and then by the Strymon Detachment (Dinter and Royden, 1993; Sokoutis et al., 1993; Brun and Sokoutis, 2007) and later steeply-dipping normal faults (Brun and Sokoutis, 2007; Wüthrich, 2009; Burg, 2011). The directions of thrusting and extension are similar, top-to-the-southwest. From the Cretaceous to ~40 Ma exhumation and the formation of detachments was related to syn-orogenic processes. From 40 Ma to Late Eocene-Early Oligocene (35-32 Ma) extension was associated with a surge of volcanic activity and no migration of volcanic centres, suggesting that the subduction zone was stationary and that extension was a consequence of lithospheric delamination, while backarc extension takes over in the Oligocene and Miocene (Wüthrich, 2009; Burg, 2011).

The internal zones of the Hellenides are bounded to the northeast by the crustal-scale NCDS that runs from offshore Mount Olympos to the northern Cyclades and further east probably in the Simav Detachment. The NCDS is originally the thrust contact between the Pelagonian domain, together with the Vardar ophiolite, and the underlying Cycladic Blueschists, that are derived from the Pindos ocean and its margins. The NCDS has been active as a post-orogenic detachment since ~32 Ma (Jolivet et al., 2010b) and the last increments of motion are as young as 8 Ma (Brichau et al., 2008; 2010; Jolivet et al., 2010b; Ring et al., 2010). Its history however started earlier, during the Eocene, when it accommodated the syn-orogenic exhumation of the H*P*-L*T* CBS. It was active contemporaneously with the basal thrust of the CBS visible on the islands of Ios and Sikinos and on top of the Menderes Massif (Huet et al., 2009). Radiometric dating shows that the WCDS and the SCDS were active throughout the Miocene (Iglseder et al., 2011; Ring et al., 2011; Grasemann et al., 2012). Motion along the WCDS may have started a little later than the NCDS, some 25 Ma ago but its activity ended approximately at the same time, some 8 Ma ago.

6.2 The Hellenic arc

In the external zones the H*P*-L*T* metamorphic units, PlattenKalk (PK) and Phyllite-Quartzite (PQ) nappes (Seidel, 1978; Seidel et al., 1982; Theye and Seidel, 1991; 1993; Brix et al., 2002), have been exhumed below the Cretan syn-orogenic Detachment (Jolivet et al., 1996; 2003) that runs from the Northern Peloponnese to the east of Crete, between the PQ nappe and the overlying Gavrovo-Tripolitza (GT) nappe that shows little metamorphic overprint. The peak of pressure, thus of burial, in the PK and PQ nappes is dated between 25 and 16 Ma (Jolivet et al., 2010c) and the detachment is associated with the formation of a supra detachment basin of Middle Miocene age (van Hinsbergen and Meulenkamp, 2006; Zachariasse et al., 2011). It is not entirely clear how much the same structure has been reactivated as a post-orogenic detachement in Crete but in the Peloponnese field evidence show a more recent reactivation on the eastern margin of the Parnon Range with shallow-dipping normal faults that extend north of the Gulf of Corinth in the Middle to the Late Miocene Itea-Amfissa detachment (Papanikolaou and Royden, 2007; Papanikolaou et al., 2009). From the Pliocene onward, extensional deformation was taken up by the steep normal faults that bounds the Parnon and Taygetos ranges in the Peloponnese or other N-S or NE-SW-trending in Crete.

6.3 The Corinth Rift

The Corinth Rift also shows a polyphased evolution. The most recent stage started some 600-700 Ka ago when the steep Helike or Xylocastro normal faults uplifted the southern margin of the rift (Armijo et al., 1996; Ford et al., 2007; Rohais et al., 2007b). These faults cut across shallow north-dipping normal faults (Sorel, 2000; Flotté et al., 2005; Jolivet et al., 2010a): the deepest is the northernmost extension of the Cretan detachment at the top of the PQ nappe and a more recent and superficial one along the contact between the GT carbonates and the underlying Tyros Beds, the Zaroukla decollement that was active before 600-700 Ka during the deposition of the early syn-rift sequence that goes back at least to 1.5 Ma and most likely to 3-4 Ma. Some shallow-dipping detachment surfaces have been imaged offshore

(Sachpazi et al., 2003; Clément et al., 2004; Taylor et al., 2011) but their relation to the onshore ones is still unclear and the amount of strain taken by these low-angle normal faults strongly debated (Bell et al., 2008; 2009; Skourtsos and Kranis, 2009). The Corinth Rift can be seen either as an equivalent of the Cyclades MCCs in an early stage of development (Jolivet, 2001; Jolivet et al., 2010a) or as a different feature due to the propagation of the NAF (Armijo et al., 1999; Taylor et al., 2011). The similarity of the crustal-scale kinematics with north-dipping detachments in both cases with comparable depths of localisation led us to prefer the first possibility, i.e. a future metamorphic core complex (Jolivet, 2001; Jolivet et al., 2010a). However a major change occurred at 600-700 Ka when the shallow-dipping detachment was cut by the now active steeply-dipping normal faults that shape the relief of the southern margin of the gulf (Rohais et al., 2007b).

6.4 Western Turkey

Western Turkey shows extensional events with a similar timing as the Aegean Sea and continental Greece. Subduction metamorphism in the Tavsanli Zone (88-80 Ma) and in the Ören-Afyon Zone (65-60 Ma) and latest Cretaceous obduction of the ophiolite over the Anatolide-Tauride block were followed by collision with the Pontides in the Early Eocene and ensuing Eocene crustal thickening episode. Extension started in the Late Oligocene – Early Miocene (ca. 30 Ma), both south of the suture in the Simav detachment and north of the suture in the Kazdag Massif and it lasted until the Middle Miocene (ca. 14-12 Ma). A series of younger detachments (Alasehir-Gediz and Buyuk Menderes detachments) then formed within the Menderes Massif (Bozkurt and Oberhänsli, 2001; Lips et al., 2001; Bozkurt and Sözbilir, 2004) and were cut at the turn of the Pliocene by the active normal faults that shape the Menderes E-W trending grabens (Seyitoglu and Scott, 1991 ; Hetzel et al., 1995a ; Emre and Sözbilir, 1997 ; Yılmaz et al., 2000; Gessner et al., 2001a ; Lips et al., 2001 ; Bozkurt and Sözbilir, 2004).

In summary: after a 20 Myrs long period of extension distributed over most of the Aegean domain, strain has localised from the Pliocene onward along the NAF and within two separate domains, in the Central Hellenic shear zone and in western Turkey. The direction of stretching and the velocity of extension have not changed significantly despite a different degree of localisation. In the earliest stages extension was accommodated by several crustal-scale detachments that reworked pre-existing shallow-dipping discontinuities in the crust, such as former thrusts. A first localisation event consisted in a faster and more intense extension in the Aegean than in Turkey with a left-lateral accommodation zone in-between in the Middle Miocene. A second event was related to the localisation of the NAF and, progressively, its south-westward extension in the Central Hellenic Shear Zone, while extension continued very actively in western Turkey.

7. Tectonic heritage and the dynamics of extension

The localisation and the evolution of structures accommodating post-orogenic extension is strongly influenced by pre-existing structures (see also (Ring et al., 2010)). The best example is the NCDS that corresponds to the reactivation of the Vardar suture zone and partly to the shear zone that exhumed the Cycladic Blueschists during the Eocene. The CBS that were buried below the Pelagonian domain were later exhumed between a detachment, or exhumation shear zone, at the top, and a thrust at the base with a mechanism that can be described as an extrusion wedge or a subduction channel. The NCDS later reactivated the exhumation shear zone. Further south, the Cretan Detachment reactivated the earlier thrust that buried the PQ nappe down to the blueschist facies conditions. Recent detachments observed in the Peloponnese also rework earlier detachments or exhumation shear zones on the eastern side of the Parnon range and south of the Corinth Rift.

The mechanical layering inherited from earlier episodes thus seems to exert a significant control on the localisation of extension and its evolution through time. This problem has been explored by means of thermo-mechanical numerical and analogue modelling (Le Pourhiet et al., 2004; Mattioni et al., 2006; Huet et al., 2011a; 2011b; Lecomte et al., 2011; 2012). It has been shown that the use of an inverted rheological stratification inherited from the nappe stacking stage with a mafic upper crust (Upper Cycladic Nappe) and a softer felsic lower crust allows the formation of MCCs with a Moho temperature much lower than classically admitted. Then, using dipping heterogeneities, also inherited from the nappe stacking episode, leads to a more asymmetric extension with several detachments all dipping in the same direction much like a large part of the Cyclades. The top-to-the-south detachments in the western Cyclades remain however unexplained by these models. Using folded contacts, both north- and south-dips could lead to a more symmetrical system or, alternatively, the strain pattern at crustal-scale becomes more asymmetrical when the amount of finite stretching increases toward the centre of the Cyclades.

8. Crustal *vs* mantle deformation

In most published numerical models the sub-lithospheric mantle behaves passively and the strain regime at the scale of the lithosphere is coaxial or, when some asymmetric asthenospheric flux is present, the consequence on the kinematics of crustal deformation are not studied in detail (Gögüs and Pysklywec, 2008a). However, a retreating subduction implies a flow of the asthenospheric mantle toward the trench (Funiciello et al., 2006) and this flow may interact with the deformation of the crust (figure 9). Getting an independent estimate of the flow pattern below the deforming Aegean lithosphere is thus an important goal. Seismic tomographic models show discontinuous slab segments along strike, suggesting a succession of tears, or gravitationally induced tensional instabilities, that have reduced its width and thus the volume of mantle impacted by the retreat. The most obvious tear has been seen in tomographic models below the transition between Anatolia and the Aegean. West of the tear the Hellenic slab dips steeper to the North than its equivalent below Turkey and an asthenospheric window has formed in-between. The inception of the STEP-fault is difficult to date precisely, but the age of the related volcanism puts it in the Middle Miocene. Although recent studies with Rayleigh waves tomography have confirmed the existence of a tear (Salaün et al., 2012), its geometry appears slightly different from a STEP-Fault as the slab seems still connected at depth. A more recent tear has been imaged below the region of the Corinth Rift and the Kephalonia Fault (Suckale et al., 2009) and Royden and Papanikolaou (2011) have proposed that it allowed the propagation of the Central Hellenic Shear Zone toward the SW until it made its junction with the Hellenic subduction zone. The age of this more recent tear can only be guessed by reference to this propagation model.

However, the Corinth Rift in its present configuration with steep normal faults is younger than 1 Ma but extension has been active in a broader zone for at least 3-4 Ma. Pliocene and Quaternary trachytic magmatic centers identified in the area of the Volos Gulf (Euboecos) and on Psathoura island have been interpreted as possibly related to the propagation of the North Anatolian Fault (Pe-Piper and Piper, 2007). Geodynamic modelling suggests an earlier initiation of the Kephalonia Fault, some 6-8 Ma ago (Royden and Papanikolaou, 2011). The third slab tear is located below the Bitlis collision zone (Faccenna et al., 2006). Tomographic models show that the slab is detached there and that the continental lithosphere is thin and hot (Piromallo and Morelli, 2003; Al-Lazki et al., 2004; Faccenna et al., 2006). Volcanism suggesting an asthenospheric window and slab delamination is recognized in the Late Miocene below eastern Turkey (Sengör et al., 2008). This would place the tear rather late, after the first one below western Turkey and before the third one below the northern Peloponnese. Faccenna et al. (2006) suggested that it may have induced the inception of the NAF. A recent investigation of marine sediments deposited on the central Anatolian plateau suggests that the slab detachment and consequential uplift (some 2 km) occurred mostly after 8 Ma (Cosentino et al., 2012).

Beside slab tears that perturbed the flow below the backarc region, gravitational forces can have induced a westward motion of the Anatolian plate along a gradient of dynamic topography. The presence of a hot and shallow asthenosphere below eastern Turkey and the dense Hellenic slab further west may be the cause of a westward flow of asthenosphere below the Anatolian plate (Faccenna and Becker, 2010; Le Pichon and Kreemer, 2010). However, it is noteworthy that the gravity push due to the dynamic topography is difficult to quantify without implementing free-surface boundary condition and rheologically realistic lithosphere in 3D convection models.

One additional possible driving force is transmitted to the Anatolian plate from the northward-moving Arabian plate across the collision zone (Armijo et al., 1999; Hubert-Ferrari et al., 2003). This mechanism is analogous to the extrusion process postulated for eastern Asia as a consequence of the collision with India (Tapponnier et al., 1982; 1986). Large-scale intracontinental strike-slip faults form as a consequence of indentation and accommodate the expulsion of rigid blocks out of the collision zone, toward free boundaries. Extrusion is usually understood as a lithospheric-scale process involving transcontinental strike-slip faults that cut down to the base of the lithosphere. The observed rigidity of the motion of the main part of Anatolia seems to fit this model in a first approach while the increase of velocity in the Aegean domain calls for a significant internal deformation that departs from a localised deformation along the northern boundary.

GPS measurements provide instantaneous displacements in the upper crust but, because of possible elastic effects due to seismic loading, it does not always give an image of the long-term (geological) flow pattern. However, a recent preliminary study of long-term crustal deformation using fault traces, fault offset rates, geodetic velocities and principal stress azimuths suggest that the Hellenic subduction is creeping and that the GPS data reflect the long-term flow pattern (Howe and Bird, 2010). A compilation of GPS data over a large region encompassing the Aegean region and the Middle East shows a large counter-clockwise circular flow with a radius of ~1200 km interpreted as the consequence of a similar asthenospheric flow about the eastern edge of the African slab, due to slab roll-back

(Le Pichon and Kreemer, 2010). SKS data below Anatolia however do not fit this pattern and the flow direction appears NE-SW under most of the Aegean and Anatolia, oblique on the E-W flow direction in the crust and on the trace of the NAF (Sandvol et al., 2003). Instantaneous upper crustal displacements (GPS), finite crustal stretching within MCCs and asthenospheric fabric (SKS) seem parallel only in the southern Aegean and in the Cyclades where the crust has been most attenuated. NE-SW fast directions are recorded as far as below the Cretan Sea suggesting that the asthenospheric flow due to slab retreat is active until the southernmost tip of the mantle wedge. The recent motion of rigid Anatolia, guided by the NAF, thus seems partly decoupled from the underlying asthenospheric flow.

Rayleigh waves anisotropy reveals the fabric of the lithospheric mantle and lower crust (Endrun et al., 2011) (figure 7). The mantle lithosphere shows a N-S fabric in the Northern Aegean roughly parallel to the asthenospheric one and to the strike of active extension in the crust, while the southern Aegean is characterized by a weak fabric with short delays matching the lack of intense recent deformation. The lower crust shows a different pattern with a strong azimuthal anisotropy up to 3.5%, parallel to the direction of stretching recorded in the Miocene MCCs of the Northern Cyclades. This pattern is interpreted as a fossilized anisotropy dating back to the Miocene (Endrun et al., 2011). Why then would the lithospheric mantle record a recent flow while the lower crust, presumably weaker, records only a fossil one ?

A different interpretation can be explored making the hypothesis that the lithosphere is thin below the central and southern Aegean in the hot backarc region leading to a weak lower crust (figures 9 and 10). The lithospheric mantle would show a fabric compatible with the N-S asthenospheric and crustal flows in the north and no significant fabric where it is thin and hot in the south. One can assume that the lithospheric mantle is too thin to induce enough seismic delays for SKS and Rayleigh waves. The upper crust has fossilized the Oligo-Miocene flow directions within the MCCs, the earliest of which having rotated clockwise while the lower crust and the uppermost mantle (Pn waves), assumed to be weak, would flow parallel to the present motion of Anatolia, parallel to the NAF. The asthenospheric flow would have remained the same from the Miocene onward and only the weak lower crust would show the recent kinematics, implying a decoupling between the asthenospheric flow and the upper crust, at the latest, after the NAF had made its junction with the Hellenic subduction zone some 5-6 Ma ago.

To summarize: SKS anisotropy suggests that the asthenospheric flow due to the retreat of the Hellenic slab is NE-SW below most of Anatolia and it rotates to a more northerly direction first progressively in Anatolia then much more abruptly in the western Aegean. The crust may have followed the same flow directions with a maximum of stretching in the Aegean region, where retreat is maximum, until 5-6 Ma. Then, the NAF made its junction with the subduction zone across the Central Hellenic Shear Zone and the crust started to move parallel to the NAF and the Central Hellenic Shear Zone. This model with a strong decoupling between the crust and the mantle explains why the asthenospheric flow does not "see" the North Anatolian Fault that is limited to the crust (in the absence of a thick lithospheric mantle).

9. Geodynamic evolution

In the following we propose a series of reconstructions (figures 10 and 11) showing the main tectonic and magmatic events in relation to the dynamics of underlying portions of slabs and mantle flow. The surface kinematics are based upon the reconstructions of Jolivet et al. (2003) and we assume a single subduction zone active since the Late Cretaceous as in Jolivet et al. (2003), Van Hinsbergen et al. (2005a), Brun and Faccenna (2008) and Jolivet and Brun (2010). This subduction zone has consumed alternatively oceanic and continental lithospheric mantle, successively the Vardar Ocean, the Pelagonian continent, the Pindos ocean, the Apulian platform and finally the eastern Mediterranean ocean. The crustal portions of those domains were progressively accreted to build the Hellenic chain and the Mediterranean Ridge accretionary complex. As our topic concerns the interactions between crustal tectonics and mantle flow during slab retreat and backarc extension, we start our reconstructions at ca. 35 Ma. On each map we show the approximate position of the slab at a depth of ~150 km as a thick light blue line and the position of magmatic centres around that time period. Symbols represent different types of magmatism after the synthetic work of Pe-Piper and Piper (2006; 2007).

35 Ma, Late Eocene (Priabonian)

This stage corresponds to the end of the subduction of the Pindos ocean in the Hellenic subduction and the end of thrusting in the Menderes. The Pelagonian domain covers the overriding plate and HP-LT units derived from the Pindos Ocean are being exhumed within the subduction channel, or extrusion wedge. Further north, a large metamorphic core complex (Rhodope) develops in the vicinity of the magmatic arc and the last compressional structures are recorded in the Balkan fold-and-thrust belt. We have represented the slab as continuous across the whole subduction system but this is a conservative option in the absence of further indications. This period is one of changing kinematic boundary conditions as the slab will start its fast southward retreat.

23 Ma, Early Miocene (Aquitanian)

Most of the Apulian lithosphere has now subducted and an accretionary wedge is developing at the expense of the external units, Gavrovo-Tripolitza, Plattenkalk and Phyllite-Quartzites. HP-LT units are exhuming within the subduction channel close to the subduction zone below the Cretan Detachment while H*T* metamorphic core complexes develop in the backarc region (Cyclades, Kazdag and northern Menderes Massifs). Slab retreat is underway and the magmatic arc has started to migrate toward the south. Asthenospheric flow is everywhere oriented toward the south or south-southwest. The exhumation of the Rhodope metamorphic core complex is still active and the NCDS (and its extension in the Simav detachment) accommodates most of the exhumation in the Cyclades and northern Menderes. A conjugate S-dipping detachment is also active in the southwestern part of the Cyclades, the West Cycladic Detachment. The subduction and migration of a single slab below the accretionary wedge induces delamination of the lower crust and mantle and all tectonic units are progressively unrooted. The replacement of the lithospheric mantle by a hot asthenosphere flowing southward induces a warmer regime in the crust and leads to the formation of H*T* metamorphic domes and partial melting of the lower crust. High-K calc-alkaline volcanic rocks are emplaced and the crustal component increases in the magmas extracted from the mantle. Crustal thickening continues in the Hellenic chain.

15 Ma, Middle Miocene (Langhian)

This stage records the last episode of exhumation of H*P*-L*T* metamorphic rocks in Crete and the Peloponnese below the Cretan Detachment. The Simav and Selale detachments in the northern Menderes and Kazdag massifs, respectively, are coming to the end of their activity. The NCDS is active in the Cyclades while thrusting proceeds in the external Hellenides. While slab retreat proceeds, the slab acquires a stronger curvature and a lateral tear forms below the western margin of Anatolia inducing an easier retreat of the western branch of the slab and a surge of alkaline volcanism. This slab tear allows the rotation of the western branch of the Hellenic belt and the Aegean. A second tear forms, probably slightly later below the Bitlis suture zone in the eastern Anatolian block. Oceanic crust of the eastern Mediterranean now sinks in the asthenosphere south of Crete.

10 Ma, Late Miocene (Tortonian)

The faster migration of the western branch of the slab in the asthenosphere decouples the Cyclades from the Menderes massif and the NCDS from the Simav Detachment. Large displacements along the NCDS are still recorded in the central Cyclades MCC, like in Mykonos and Naxos, contemporaneously with the intrusion of I-type plutons. Crustal thinning has now reached the Cretan Sea and the Central Hellenic Shear Zone starts to form as a western prolongation of the North Anatolian Fault above a flow of mantle toward the south. The Mediterranean Ridge accretionary prism progressively develops south of Crete at the expense of sediments carried by the subducting eastern Mediterranean oceanic crust.

5 Ma, Pliocene (Zanclean) to Present

While the slab continues its southward retreat, inducing a southward asthenospheric flow, a third tear forms within slab below the present Corinth Rift. The Aegean portion of the retreating slab is now decoupled from the main branch to the west and migrates southward or southwestward. This new situation facilitates the junction between the North Anatolian Fault and the Hellenic Trench through the Central Hellenic Shear Zone and the Kephalonia Fault. Guided by the NAF to the north, Anatolia moves faster westward as a rigid block and active deformation progressively localises along the NAF and its junction with the trench as well as in the western Anatolian grabens. The exact starting date of the fast motion of Anatolia is difficult to assess, the velocity increased probably between 10 and 5 Ma with a localisation of the NAF in the Dardanelles some 6 Ma ago. During this period the motion of Anatolia is decoupled from the underlying asthenospheric flow and motion vectors are parallel to the direction of slab retreat only in the southern Aegean.

10. Discussion

10.1 Slab fragmentation and strain localisation in the Aegean

Since 35 Ma slab retreat has controlled crustal deformation in the Aegean. While exhumation of H*P*-L*T* metamorphic units proceeded within the subduction interface, in an episodic fashion, or more or less continuously, backarc extension has led to the formation of MCCs topped by a small number of major detachments such as the NCDS or the WCDS, above a southward flowing asthenosphere that replaces the lithospheric mantle and induces a high heat flow and a weak crust in the Cyclades and Menderes. The NCDS, as well as the Cretan detachments, is inherited from the Eocene crustal thickening episode. The geological record shows a stepwise localisation of strain until the present situation where the NAF accomodates the westward displacement of Anatolia, and the Corinth Rift and the grabens of Western Turkey take most of the extension in the backarc. A first significant event is the decoupling between the Aegean and the Menderes above a tear in the slab in the Middle Miocene. A second tear happens in the east of Anatolia at a date that is difficult to exactly define but which is between 15 and 10 Ma. A third tear is probable in the Late Miocene below the future Corinth Rift, allowing the junction between the NAF and the trench and the fast westward motion of Anatolia. At this stage the crustal motion is mostly imposed by the geometry of the NAF in northern Turkey and it is highly oblique to the asthenospheric flow below.

The first order observation that most recent and active strike-slip faults do not cross the NCDS and do not propagate in the extensional domain (neither in the Cyclades nor in the regions of grabens between Evvia and the Corinth Rift) suggests a different rheological behaviour north and south of the NCDS. The two regions are indeed characterized by different mechanical stratification, with a greater importance of the Apulian domain and its ophiolite in the north and a thicker partially molten crust in the south. The crustal thickness is also different with a thinner crust in the north which may contribute to smaller gravitational forces.

A classical concept is that the propagation of the NAF induced crustal scale tension gashes at its termination and that the most recent one is the Corinth Rift. GPS data show that the whole Aegean domain and the whole Peloponnesus are moving coherently faster than the northern Aegean, implying a major component of traction along the southern limit. Much of the retreat was achieved before the formation of the Gulf of Corinth and deformation was until then much less localised. A second slab tearing to the west of the retreating domain could have enabled the continuation of retreat with a more rigid slab retreating between two tears, one below western Turkey and one below the Corinth Rift and Kephalonia Fault. The presence of the two tears rendered unnecessary a strong deformation of the slab and thus the crustal block above could move coherently southward, which can explain the absence of deformation in the Cyclades during the recent period.

The tear below the Peloponnesus is the last stage of a long deformation history of the Hellenic slab, leading to the isolation of a narrow stripe of subducting lithosphere that can move backward easily in the mantle (Suckale et al., 2009; Royden and Papanikolaou, 2011). A similar evolution is shown in the Western and Central Mediterranean where an initially long slab from the Rif-Betics to the Northern Apennines has been progressively reduced to only two narrow stripes during the retreat, one below the Gibraltar Arc and one below Sicily (Faccenna et al., 2004; Spakman and Wortel, 2004; Jolivet et al., 2008). Backarc extension started when the subduction regime changed at the scale of the Mediterranean some 30-35 Ma ago. Then a first tear happened below western Turkey allowing a large retreat of the slab inducing a rotation of continental Greece about a pole located in NW Greece or Albania. The last tear below Corinth has finally totally broken the slab and the retreat process was facilitated. During the same period a slab breakoff occurred to the east below the Bitlis suture that could have been the cause of the initiation of the North Anatolian Fault and the westward motion of Anatolia. Before the slab was broken below the Peloponnesus, the velocity of extrusion was limited by the velocity of slab retreat. Since the slab has been torn there the velocity of extrusion has increased and a plate limit has formed to the west, leading to the connection of the NAF with the Kephalonia Fault through a series of active grabens such as the Corinth Rift. Because of the gravitational forces created by the thickened Peloponnese crust, extensional shear zone predominates over strike-slip fault within the dextral step over formed between the Kephalonia fault and the NAF.

10.2 The North Anatolian Fault and the Aegean

In this hypothetical model we see two tectonic behaviours. Most of the history of backarc extension is driven from below by slab retreat but from the Late Miocene the displacement of Anatolia is decoupled from the underlying mantle flow. A simple approach is to associate the westward motion of Anatolia and the Arabia-Eurasia collision through an extrusion model. In this case, the Anatolian plate would be pushed by the Arabian plate across the Bitlis suture zone and most of the stress would be transmitted through the upper crust as the lithospheric mantle appears very thin below most of Anatolia. The tectonic evolution of

Anatolia since the formation of the NAF would then be driven from above. This would fit the difference between the stretching direction in the mantle (SKS) and the crustal flow shown by GPS measurements. It would also fit the rigid behaviour of Anatolia. Extrusion, that is usually understood as a lithospheric scale process would then be here restricted to the upper crust. An alternative vision is to consider that the westward motion of Anatolia is simply carried by a westward mantle flow due to delamination and an asthenospheric plume that would not yet be recorded in the SKS fabric because it is too recent. The exact timing of the formation of the NAF and its propagation, compared to the timing of the delamination (constrained by the age of volcanim) is a crucial question that would help answering this question.

10.3 Crust-mantle interactions during slab retreat

In this paper we suggest that the asthenospheric flow controls the deformation in the upper crust. How is this control then achieved ? One possibility is that the asthenospheric flow imposes at the base of the lithosphere a drag strong enough to control crustal deformation. This supposes that the lithospheric mantle and the lower crust are as weak as the asthenosphere, so that the flow is homogeneous in the whole column below the upper crust. This corresponds to the Aegean situation where the crust was thick and the lithospheric mantle thin when extension started. In most subductions zones the backarc and the fore-arc domains move toward the trench (Faccenna et al., 2007; Heuret et al., 2007) suggesting a role played by such a basal drag. A right combination of a high strain rate imposed by the asthenospheric flow on a large domain and a low viscosity may lead to an efficient coupling between asthenospheric flow and crustal deformation. The hypothesis of a control by the basal drag imposed by asthenospheric flow is thus to be explored for a crust no more resistant on the long term (geological durations) than its weakest low-angle normal fault for instance.

An alternative is to consider that slab retreat induces 3D mantle currents that in turn create a dynamic topography with gradients of potential energy along which continental blocks can move. This 3D flow would result from a gradient of dynamic topography between an upwelling of asthenosphere below the collision zone and the downgoing slab in the Aegean. Using the distribution of temperature anomalies derived from seismic tomographic models, Faccenna and Becker (2010) have modelled flow directions in the mantle of the Mediterranean region. They reproduce the westward motion of Anatolia but the model fails to account for the southward motion of the Aegean slab and the N-S extension in the backarc region, and more generally the toroidal flow suggested at a larger scale in GPS data

(Le Pichon and Kreemer, 2010). It can be argued that temperature anomalies cannot be safely recovered from seismic velocity anomalies without good knowledge of mantle rheology and composition (that affect anharmonic and anelastic contribution to seismic velocities).

Both hypotheses would fit the observed similarity between the SKS fast directions and the stretching lineations in the crust. Then, models involving dynamic topography should explain not only the westward movement of Anatolia (Faccenna and Becker, 2010) but also the gradient of extension from east to west with the westernmost part of the system spreading faster, as they should explain why extension continues after the continental lithosphere has been attenuated and replaced by oceanic crust as in the Liguro-Provençal Basin and Tyrrhenian Sea for instance. This question is thus open and it should be a goal for future numerical studies. We finally favor a model involving (1) slab retreat and related asthenospheric flow that control the strain within regions where the heterogeneous crust is thin and weak and the mantle hot, and (2) some rigid blocks at the free surface, able to transmit stresses horizontally over large distances. Such a model may better account for the observations but it remains to be tested numerically.

11. Conclusions

The Aegean region shows a clear localisation of deformation through time. But if we come back to the initial debate between an intrinsically localising continental lithosphere leading to the formation of large-scale strike-slip faults and a overall weak lithosphere that would distribute the strain on large domains, the question has significantly changed.

The first point is that the continental crust has inherited a strong heterogeneity from its earlier tectonic history and that crustal-scale contacts such as major thrust planes act (1) as weak zones that can localise later deformation during exhumation and backarc extension and (2) as zones of strong contrasts of resistance and viscosity that also have a significant influence on strain localisation and the kinematics of extension. Moreover, the succession of a mountain building in the Eocene and a backarc extension in the Oligo-Miocene has left the Aegean with different crusts north and south of the NCDS. North of it the Pelagonian domain and its ophiolites is present and the crust is thin, while south of it the Pelagonian domain is represented only as small klippes and H*T* domes have been exhumed.

The second point concerns the coupling between mantle and crustal processes. The dynamics of slabs and slab portions at depth and the asthenospheric flow due to slab retreat has a major influence on the localisation of deformation in the upper plate. Successive slab ruptures along the strike of the Hellenides and Taurides from the Middle Miocene to the Late Miocene have progressively isolated a narrow stripe of lithosphere, still attached to the African lithosphere at the longitude of Crete. The formation of the North Anatolian Fault and its propagation within the Central Hellenic Shear Zone is partly a consequence of this evolution. Once the connection of the NAF with the trench through the CHSZ has been made the extrusion of Anatolia seems to have reoriented the flow of the upper crust and the strain in the lower crust. The extrusion of Anatolia and the Aegean extension are thus partly driven from below (asthenospheric flow) and from above (extrusion of a lid of rigid crust).

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Figure captions

Figure 1 : A series of maps showing the main structures and data sets discussed in the text.

A: Tectonic map of the Aegean and Anatolian region showing the main active structures (black lines), the main sutures zones (thick violet or blue lines), the main thrusts in the Hellenides where they have not been reworked by later extension (thin blue lines), the North Cycladic Detachment (NCDS, in red) and its extension in the Simav Detachment (SD), the main metamorphic units and their contacts; AlW : Almyropotamos window ; BD : Bey Daglari ; CB : Cycladic Basement ; CBBT : Cycladic Basement basal thrust ; CBS : Cycladic Blueschists ; CHSZ : Central Hellenic Shear Zone ; CR : Corinth Rift ; CRMC : Central Rhodope Metamorphic Complex ; GT : Gavrovo-Tripolitza Nappe ; KD : Kazdag dome ; KeD : Kerdylion Detachment ; KKD : Kesebir-Kardamos dome ; KT : Kephalonia Transform Fault ; LN : Lycian Nappes ; LNBT : Lycian Nappes Basal Thrust ; MCC : Metamorphic Core Complex ; MG : Menderes Grabens ; NAT : North Aegean Trough ; NCDS : North Cycladic Detachment System ; NSZ : Nestos Shear Zone ; OlW : Olympos Window ; OsW : Ossa Window ; OSZ : Ören Shear Zone ; Pel. : Peloponnese ; ÖU : Ören Unit ; PQN : Phyllite-Quartzite Nappe ; SiD : Simav Detachment ; SRCC : South Rhodope Core Complex ; StD : Strymon Detachment ; WCDS : West Cycladic Detachment System ; ZD : Zaroukla Detachment.

B: Seismicity. Earthquakes are taken from the USGS-NEIC database. Color of symbols gives the depth (blue for shallow depths) and size gives the magnitude (from 4.5 to 7.6).

C: GPS velocity field with a fixed Eurasia after Reilinger et al. (2010).

D: the domain affected by distributed post-orogenic extension in the Oligocene and the Miocene and the streching lineations in the exhumed metamorphic complexes.

E: The thick blue lines illustrate the schematized position of the slab at ~150 km according to the tomographic model of Piromallo and Morelli (2003), and show the disruption of the slab at three positions and possible ages of these tears discussed in the text. Velocity anomalies are displayed in percentages with respect to the reference model sp6 (Morelli and Dziewonski, 1993). Colored symbols represent the volcanic centres between 0 and 3 Ma after Pe-Piper and Piper (2006).

F: Seismic anisotropy obtained from SKS waves (blue bars, (Paul et al., 2010)) and Rayleigh waves (green and orange bars, Endrun et al., 2010). See also (Sandvol et al., 2003). Blue lines show the direction of stretching in the asthenosphere, green bars represent the stretching in the lithospheric mantle and orange bars in the lower crust.

G: Focal mechanisms of earthquakes over the Aegean Anatolian region. Source: (CMT, 2011). Base maps made with GeoMapApp (http://www.geomapapp.org)(Ryan et al., 2009).

Figure 2 : Detail of focal mechanisms of earthquakes of magnitude greater than 4. Source: NEIC 200-2011 and (Kiratzi and Louvari, 2003). Green: strike-slip, blue: normal, red: reverse faults.

Figure 3: Compiled tectonic map of the Aegean region, Menderes massif, Rhodope massif and the Balkan, adapted from (Jolivet and Brun, 2010), after (Creutzburg, 1977; Bonneau, 1982; Bonneau and Kienast, 1982; Bonneau, 1984; Lyon-Caen et al., 1988; Burg et al., 1990; Armijo et al., 1992; Burg et al., 1995; Burg et al., 1996; Tzankov et al., 1996; Collins and Robertson, 1997; Armijo et al., 1999; Collins and Robertson, 1999; Okay and Tüysüz, 1999; Okay and Satir, 2000; Koukouvelas and Aydin, 2002; Jolivet et al., 2004b; Papanikolaou et al., 2004; Chamot-Rooke et al., 2005) and the Geological map of Greece (IGME). Moho depth is after Tirel *et al.* (2004b).

Figure 4 : A N-S cross-section of the Aegean region from the northern passive margin of Africa to the Moesian platform across Crete, the Cyclades and the Rhodope massif after Jolivet and Brun (2010) and the southward migration of metamorphic, tectonic and magmatic processes.

Figure 5 : Maps showing the stretching lineations and kinematic indicators separated by age (left: Eocene and pre-Eocene; right: Oligocene and Miocene) and by tectonic context (blue arrows: thrust-related deformation, green arrows: syn-orogenic exhumation-related deformation, red arrows: post-orogenic extension-related deformation. Base maps made with GeoMapApp (http://www.geomapapp.org)(Ryan et al., 2009)

Figure 6 : Horizontal sections through the tomographic model of Piromallo and Morelli (2003). The white arrows show the three possible tears in the slab.

Figure 7 : Comparison of the directions of stretching lineations and the directions of stretching obtained from seismic anisotropy in the Aegean region (left: asthenosphere –SKS waves –red and orange bars from Paul et al., 2010 and Hatzfeld et al. 2001, blue circles for null anisotropy, middle: lithospheric mantle –Rayleigh waves-, right: lower crust –Rayleigh waves-, lower: uppermost mantle –Pn waves-). Base maps made with GeoMapApp (http://www.geomapapp.org)(Ryan et al., 2009).

Figure 8: The differents types of magmatic products in the Aegean region, redrawn after Pe-Piper and Piper (Pe-Piper and Piper, 2007).

Figure 9 : Three reconstructions of the section of figure 3 showing the progressive slab retreat after Jolivet and Brun (2010) and a velocity field of particles after the analogue model of Funiciello et al. (Funiciello et al., 2003). Partially molten lower crust is shown in red.

Figure 10 : Two tentative 3D reconstructions and flow directions in the mantle (blue arrows) and upper (red arrows) and lower (orange arrows) crusts of the Aegean region before the recent slab tear below the Corinth Rift and after.

Figure 11 : Reconstructions of the Aegean region from the Late Eocene (35 Ma) to the Present. The thick blue line shows the position of the slab at a depth of 150 km. The blue domain is the oceanic lithosphere of the eastern Mediterranean. Green arrows represent the asthenospheric flow and orange arrows the upper crustal flow. Volcanism from Pe-Piper and Piper (2006; 2007).

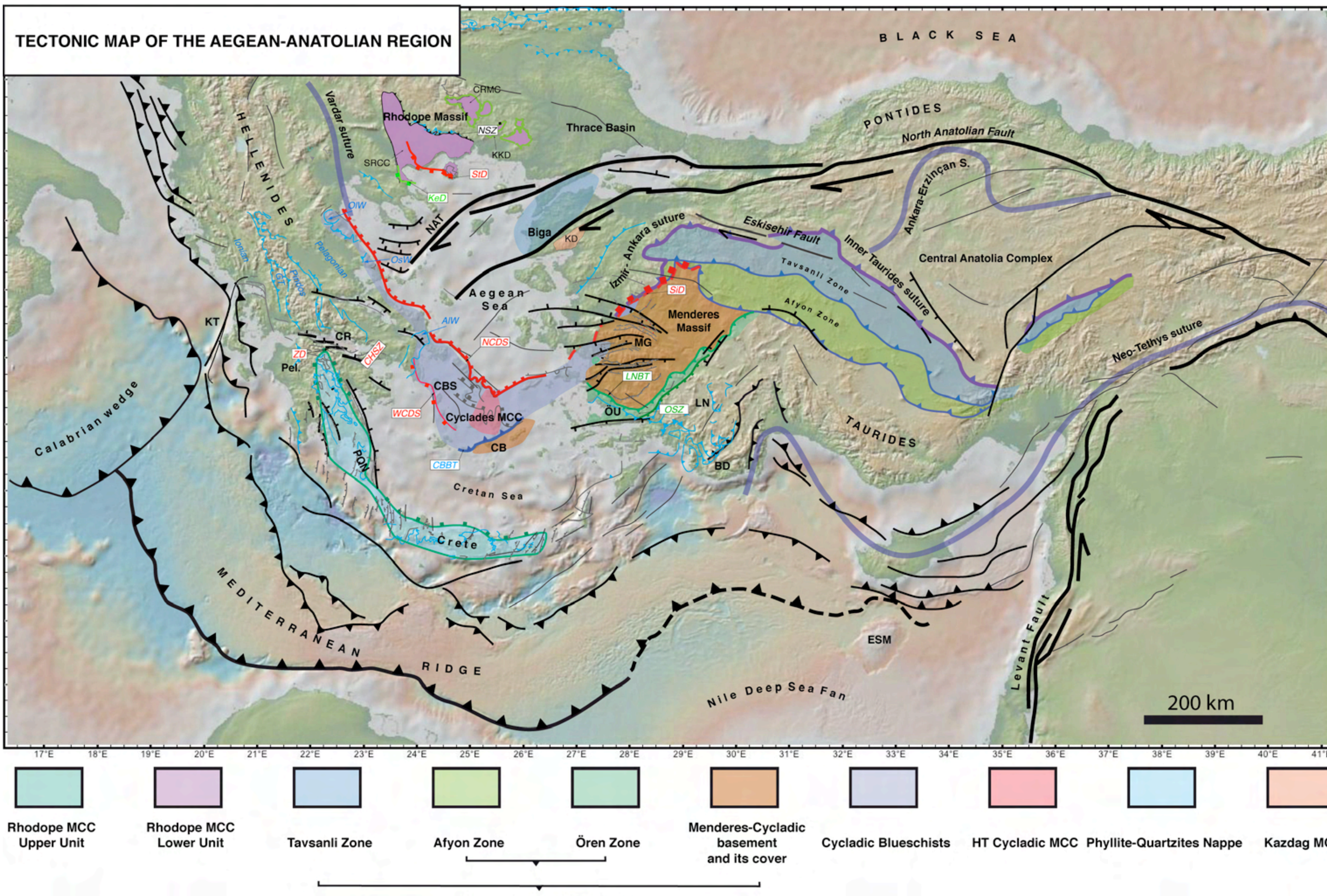
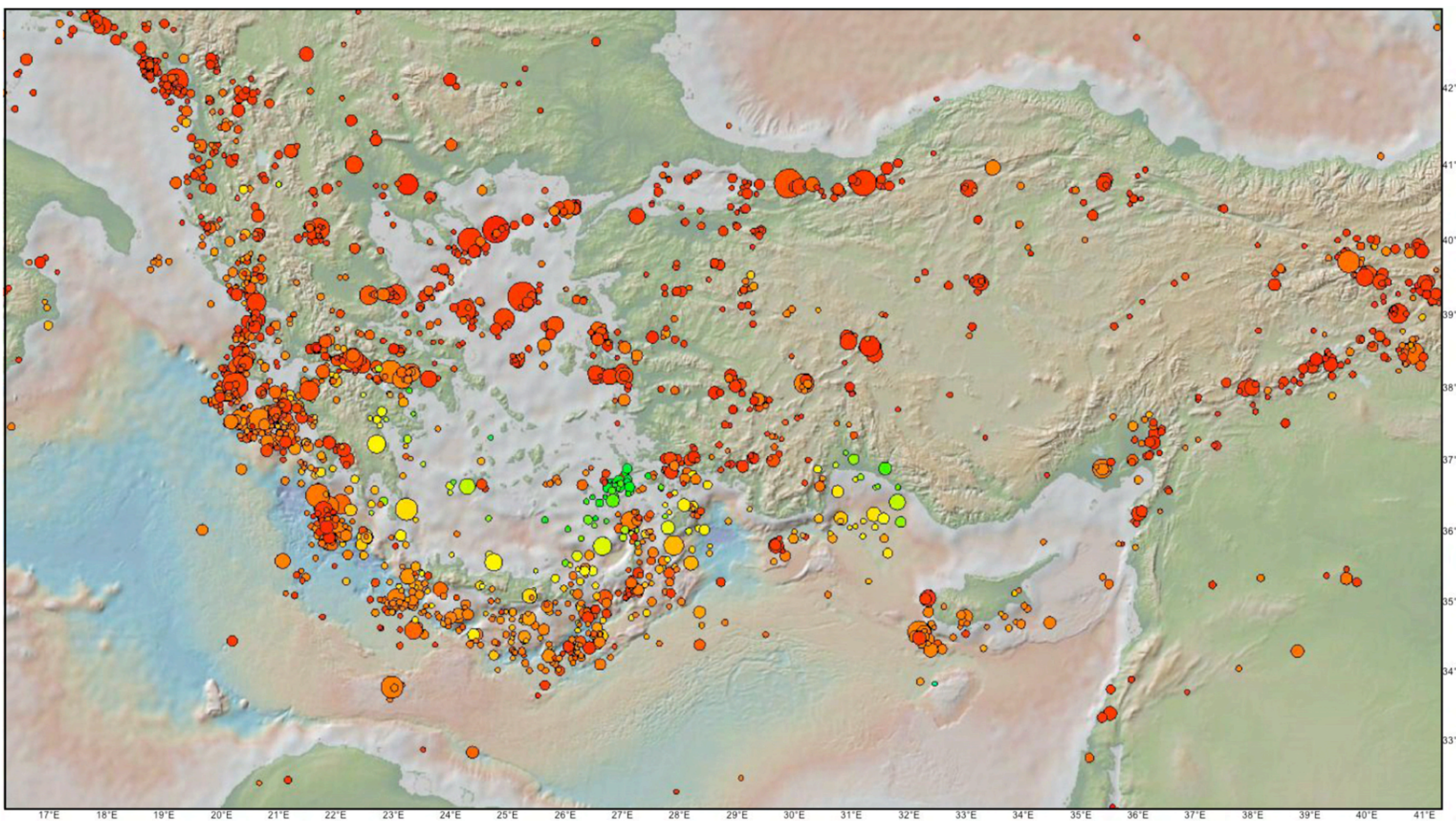

TECTONIC MAP OF THE AEGEAN-ANATOLIAN REGION
BLACK SEA
PONTIDES
North Anatolian Fault
Rhodope Massif
Thrace Basin
Vardar suture
HELLENIDES
NAT
Biga
KD
Aegean Sea
Izmir - Ankara suture
Eskisehir Fault
Inner Taurides suture
Ankara-Erzinçan S.
Central Anatolia Complex
Tavsanli Zone
Afyon Zone
Menderes Massif
MG
LNBT
ÖU
OSZ
LN
BD
SiD
CBS
Cyclades MCC
CB
NCDS
WCDS
CBBT
AIW
CHSZ
ZD
CR
Pel.
KT
PQN
Crete
Cretan Sea
TAURIDES
Neo-Tethys suture
Calabrian wedge
MEDITERRANEAN RIDGE
Nile Deep Sea Fan
ESM
Levant Fault
200 km
42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E
Rhodope MCC Upper Unit
Rhodope MCC Lower Unit
Tavsanli Zone
Afyon Zone
Ören Zone
Menderes-Cycladic basement and its cover
Cycladic Blueschists
HT Cycladic MCC
Phyllite-Quartzites Nappe
Kazdag MCC
Anatolides-Taurides Block

Figure 1A



42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E

Figure 1B

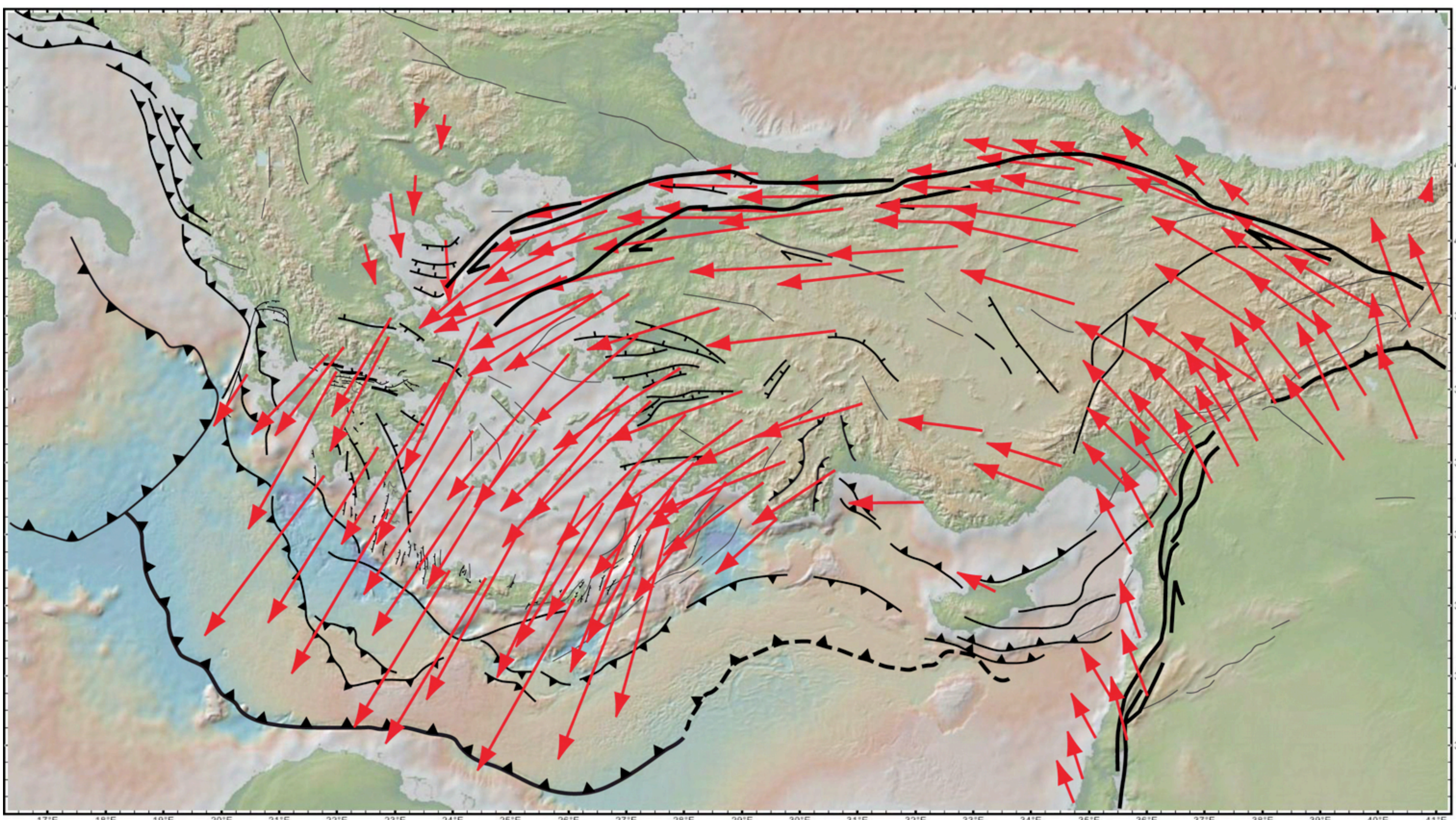
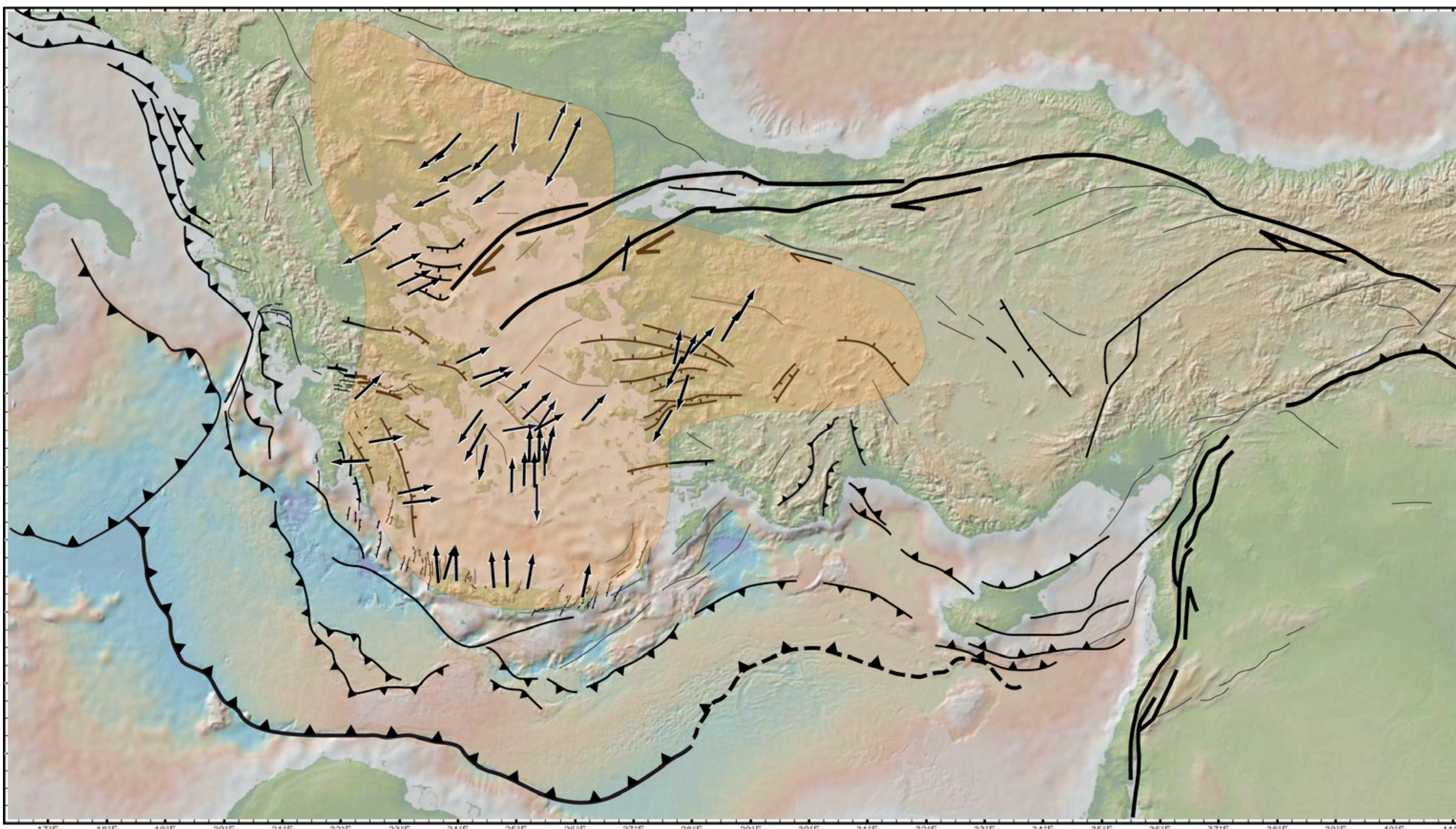

42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E

Figure 1C



42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E

Figure 1D

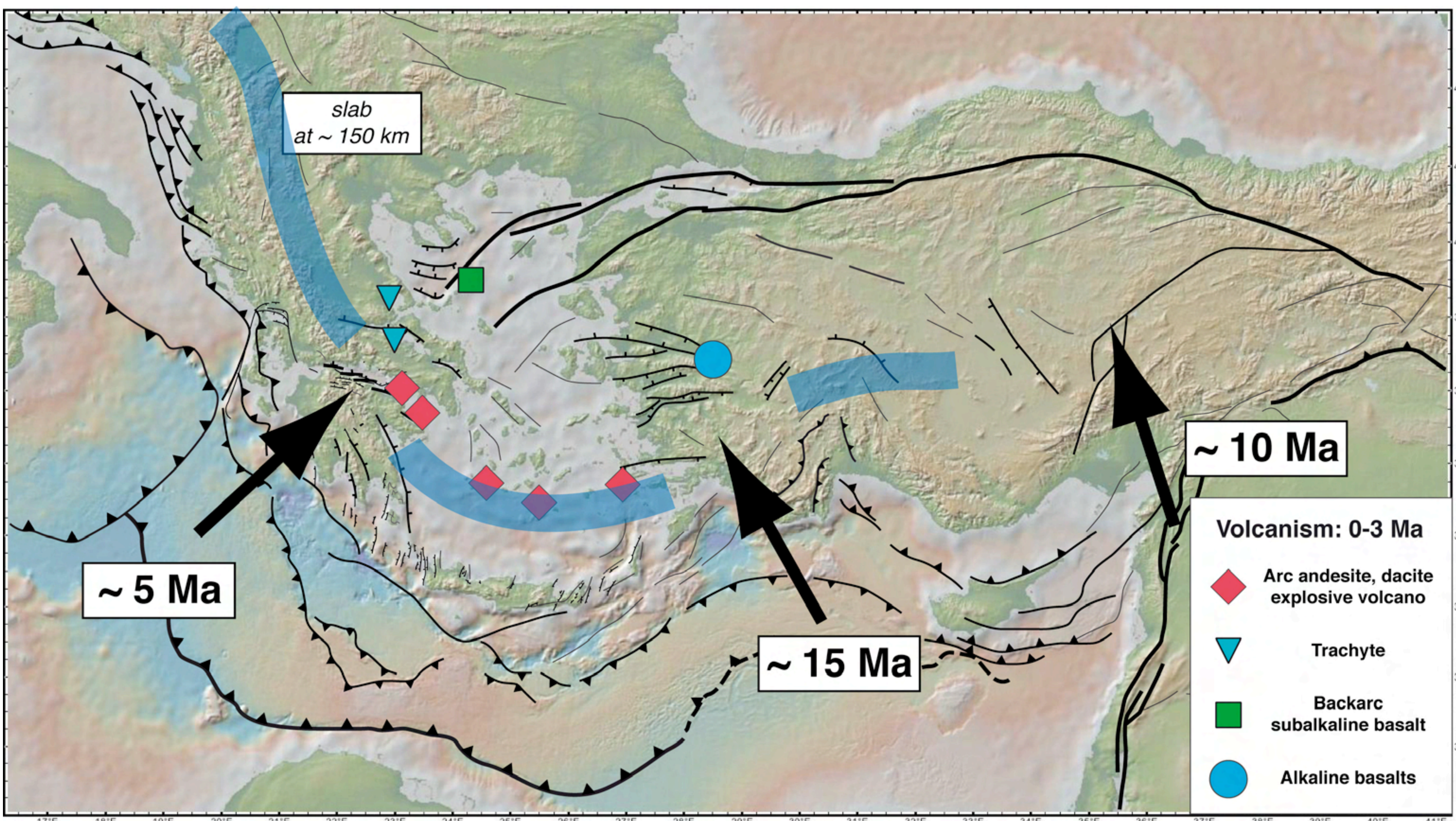

slab
at ~ 150 km
~ 5 Ma
~ 15 Ma
~ 10 Ma
Volcanism: 0-3 Ma
Arc andesite, dacite explosive volcano
Trachyte
Backarc subalkaline basalt
Alkaline basalts

Figure 1E

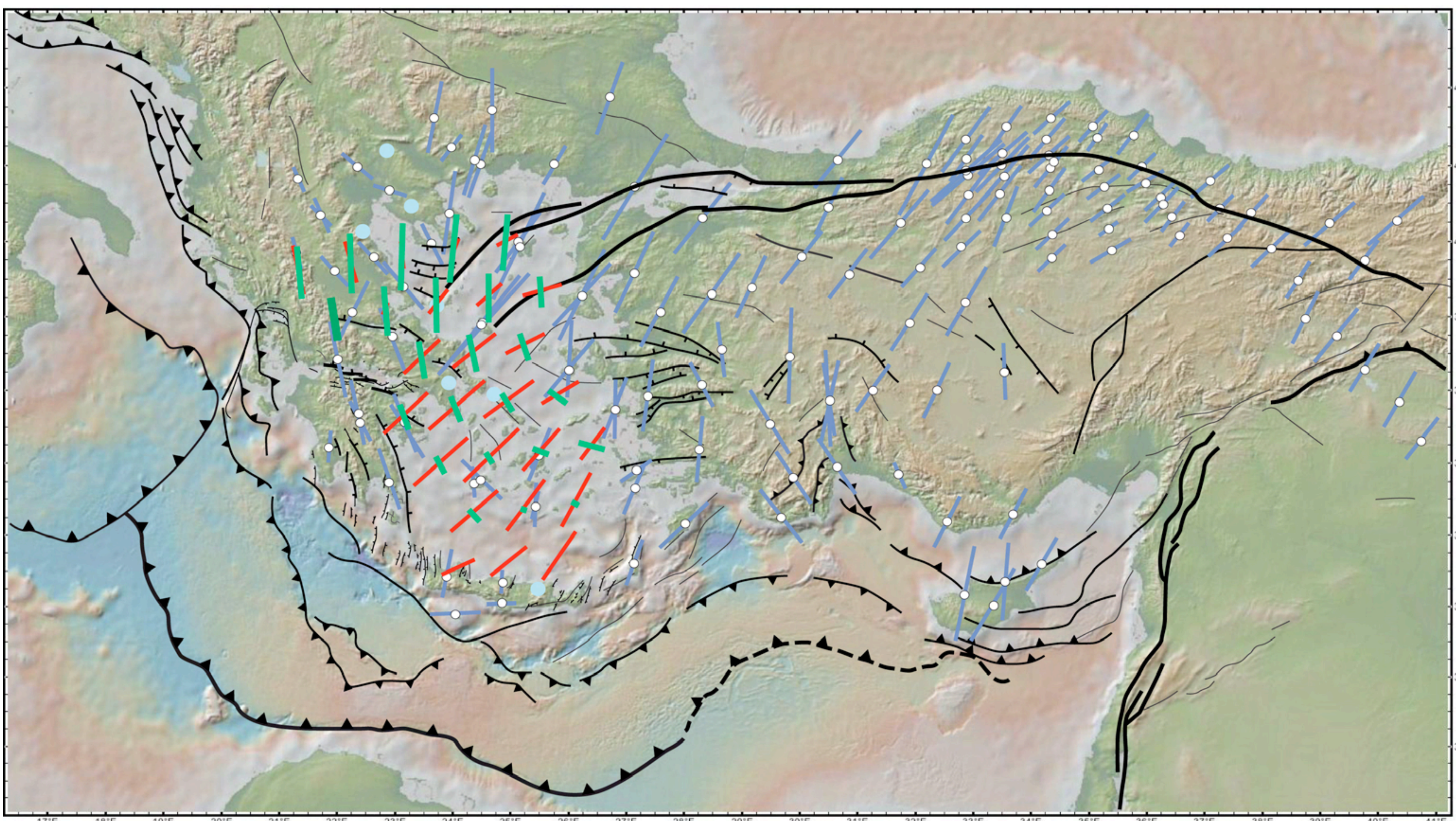

42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E

Figure 1F

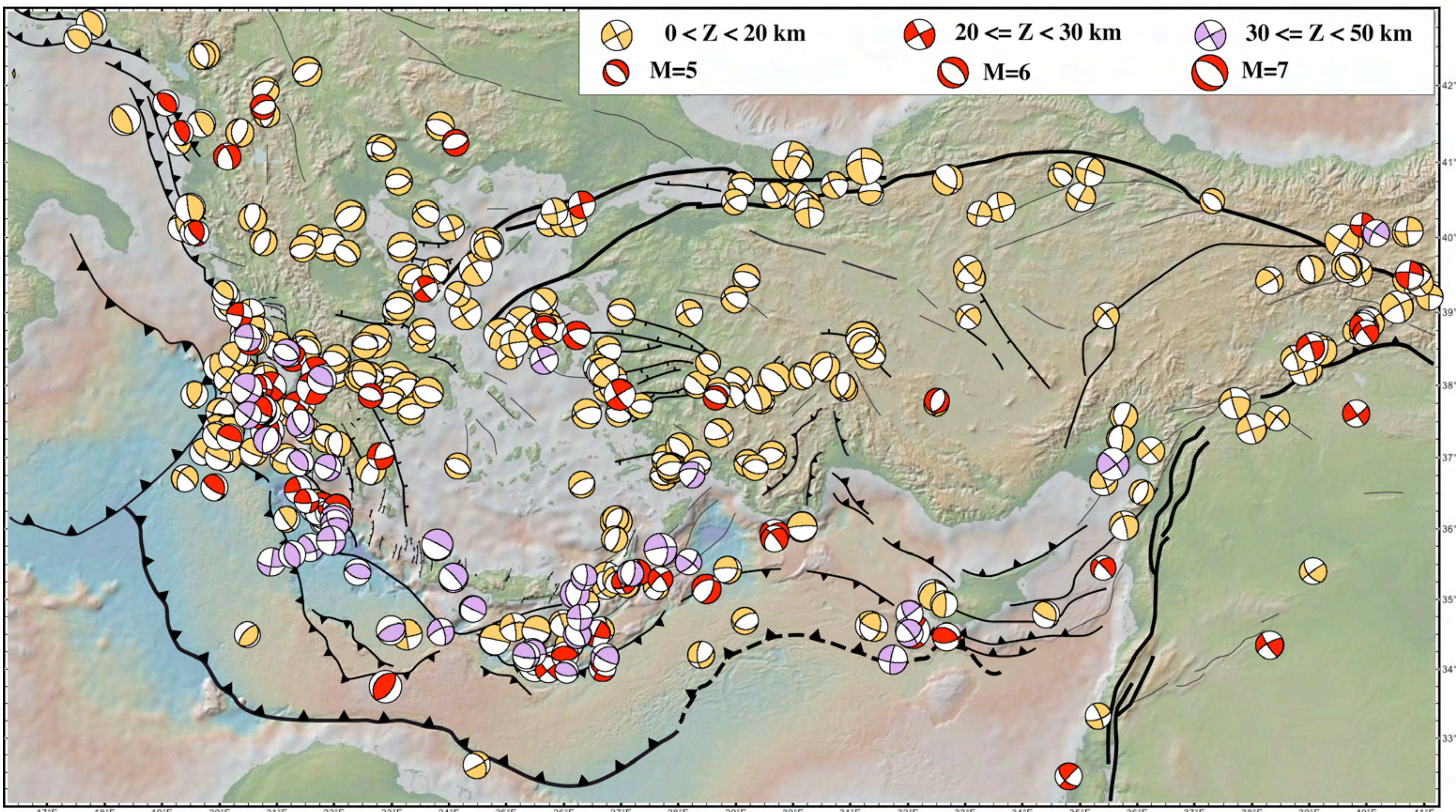

0 < Z < 20 km
20 <= Z < 30 km
30 <= Z < 50 km
M=5
M=6
M=7
42°N
41°N
40°N
39°N
38°N
37°N
36°N
35°N
34°N
33°N
17°E
18°E
19°E
20°E
21°E
22°E
23°E
24°E
25°E
26°E
27°E
28°E
29°E
30°E
31°E
32°E
33°E
34°E
35°E
36°E
37°E
38°E
39°E
40°E
41°E

Figure 1G

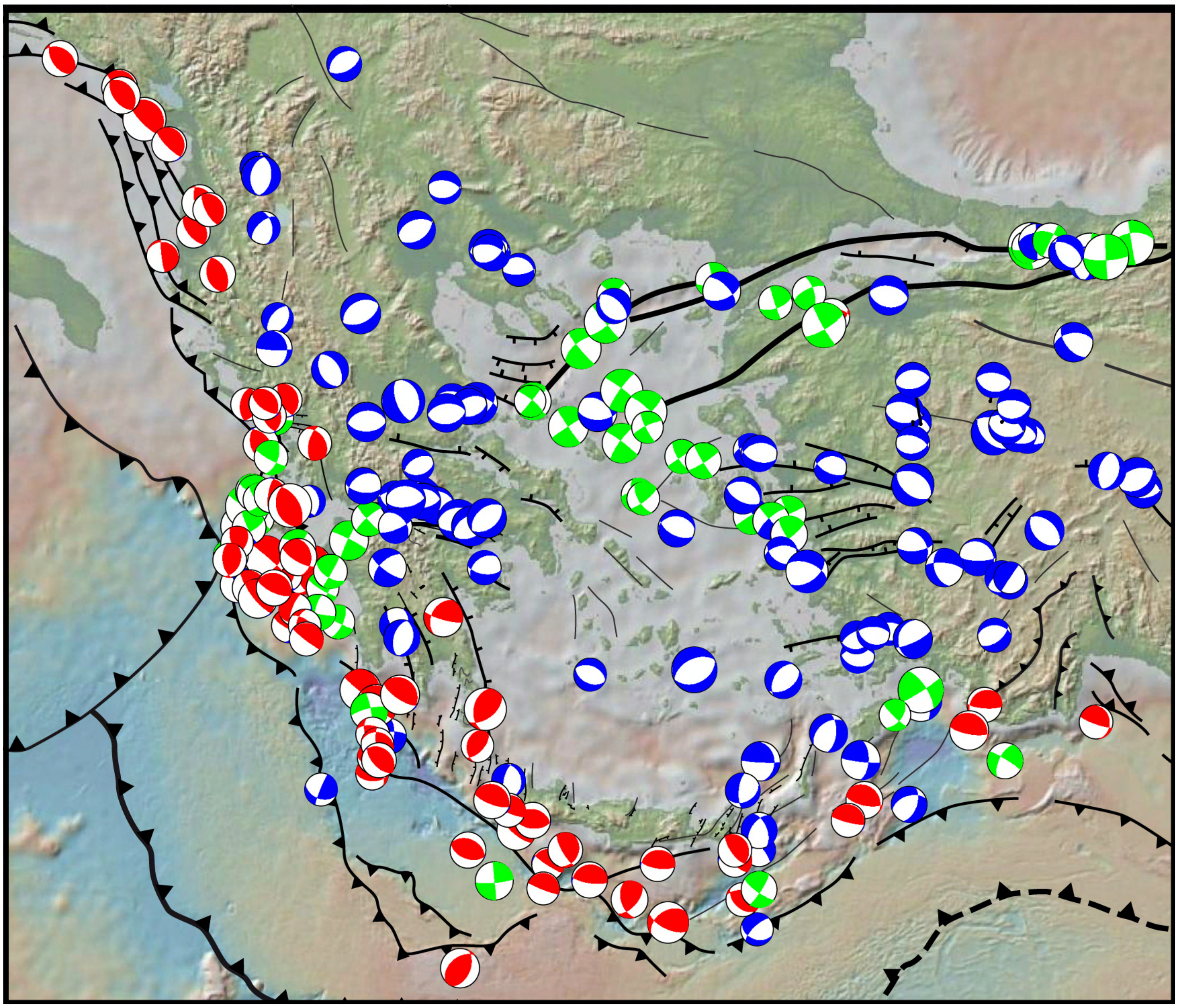


Figure 2

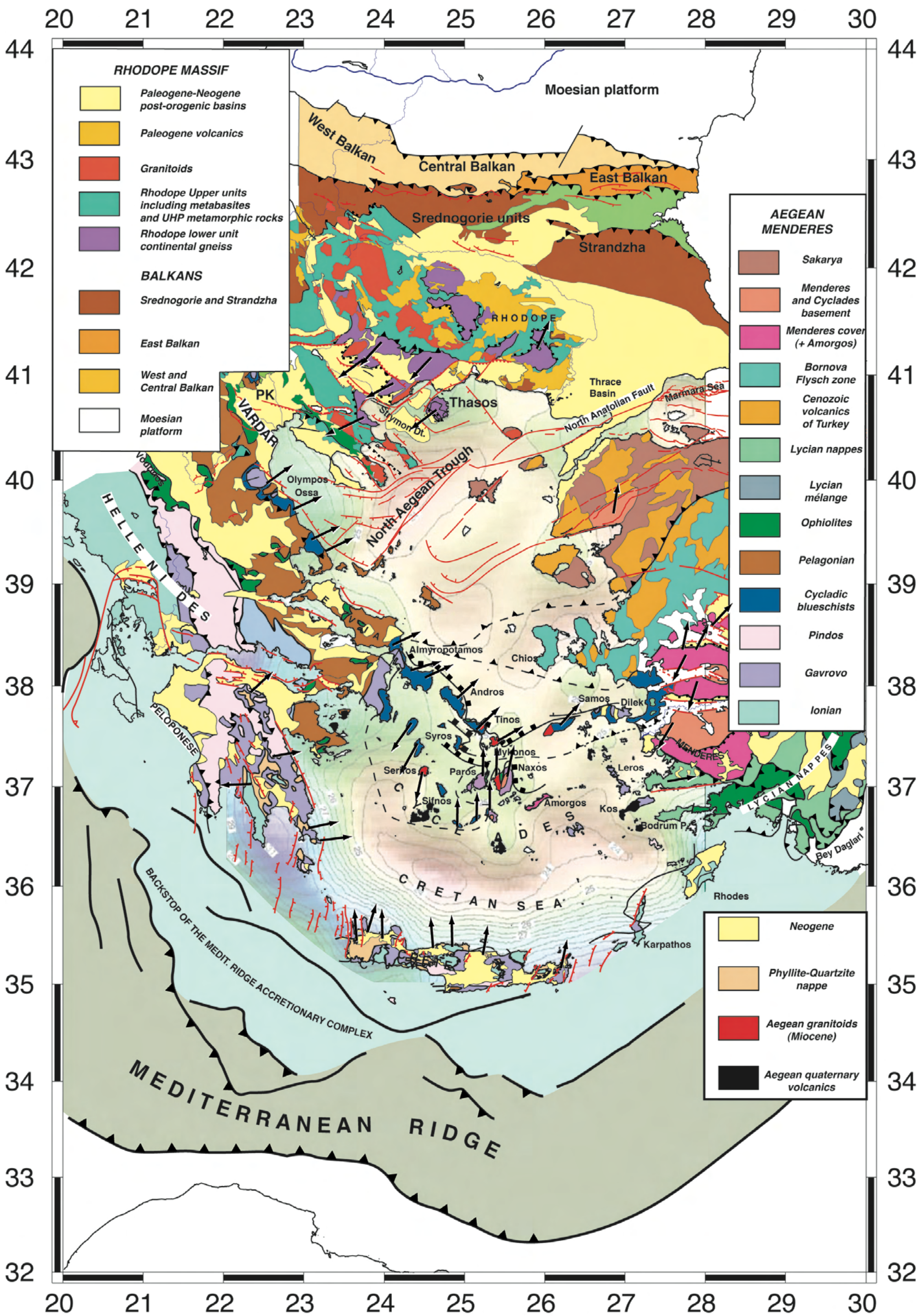

RHODOPE MASSIF
Paleogene-Neogene post-orogenic basins
Paleogene volcanics
Granitoids
Rhodope Upper units including metabasites and UHP metamorphic rocks
Rhodope lower unit continental gneiss
BALKANS
Srednogorie and Strandzha
East Balkan
West and Central Balkan
Moesian platform
AEGEAN MENDERES
Sakarya
Menderes and Cyclades basement
Menderes cover (+ Amorgos)
Bornova Flysch zone
Cenozoic volcanics of Turkey
Lycian nappes
Lycian mélange
Ophiolites
Pelagonian
Cycladic blueschists
Pindos
Gavrovo
Ionian
Neogene
Phyllite-Quartzite nappe
Aegean granitoids (Miocene)
Aegean quaternary volcanics
Moesian platform
West Balkan
Central Balkan
East Balkan
Srednogorie units
Strandzha
RHODOPE
Thrace Basin
North Anatolian Fault
Marmara Sea
PK
VARDAR
Thasos
Strymon Dt.
Olympos
Ossa
North Aegean Trough
HELLENIDES
Almyropotamos
Chios
Andros
Samos
Dilek
Tinos
Syros
Mykonos
Naxos
Serifos
Paros
Sifnos
Amorgos
Leros
Kos
Bodrum P.
CYCLADES
MENDERES
LYCIAN NAPPES
Bey Daglari
PELOPONESE
CRETAN SEA
Rhodes
Karpathos
BACKSTOP OF THE MEDIT. RIDGE ACCRETIONARY COMPLEX
MEDITERRANEAN RIDGE
20
21
22
23
24
25
26
27
28
29
30
44
43
42
41
40
39
38
37
36
35
34
33
32

Figure 3

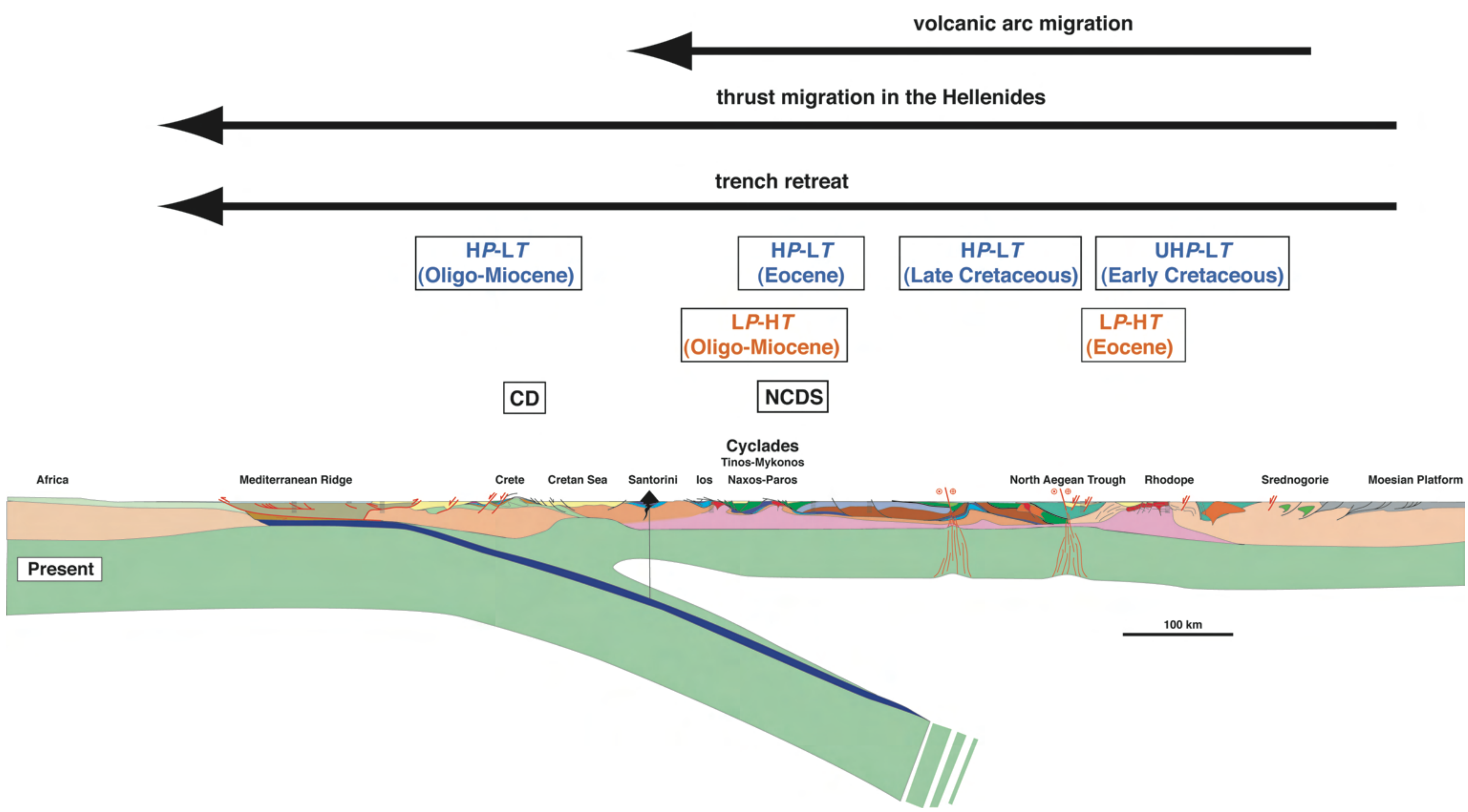
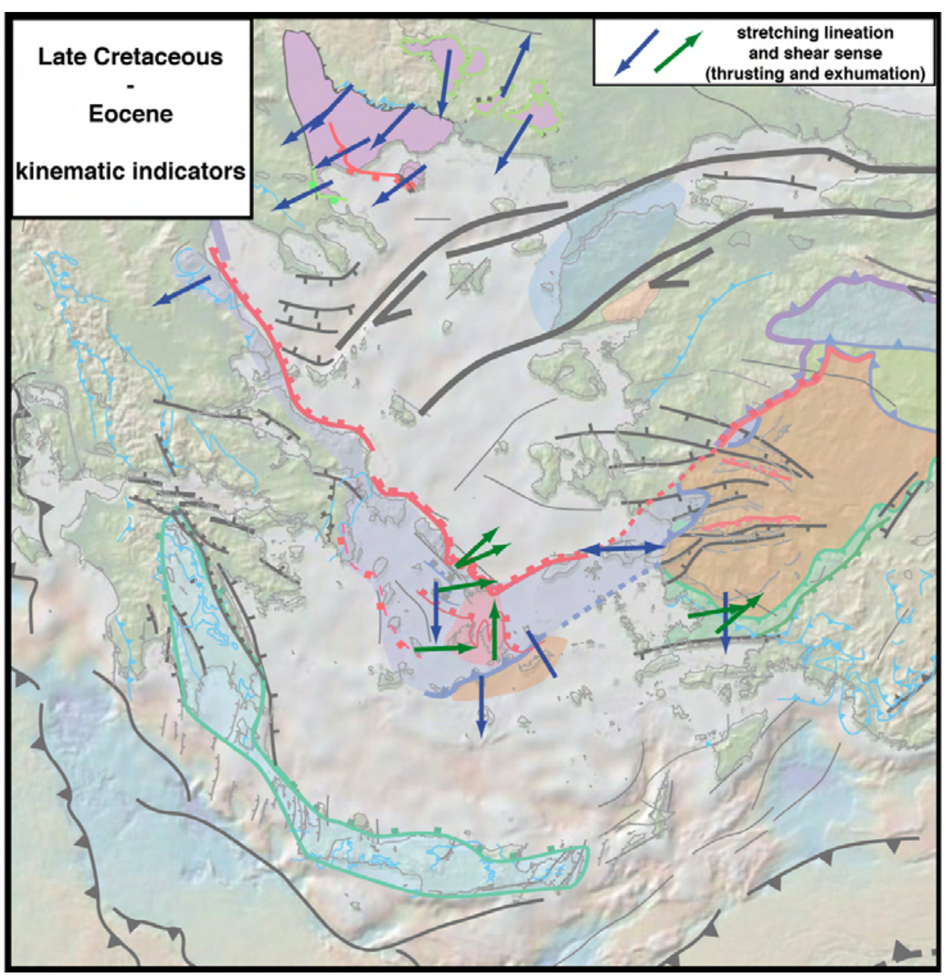

volcanic arc migration
thrust migration in the Hellenides
trench retreat
HP-LT
(Oligo-Miocene)
HP-LT
(Eocene)
HP-LT
(Late Cretaceous)
UHP-LT
(Early Cretaceous)
LP-HT
(Oligo-Miocene)
LP-HT
(Eocene)
CD
NCDS
Cyclades
Tinos-Mykonos
Africa
Mediterranean Ridge
Crete
Cretan Sea
Santorini
Ios
Naxos-Paros
North Aegean Trough
Rhodope
Srednogorie
Moesian Platform
Present
100 km

Figure 4


Late Cretaceous
-
Eocene
kinematic indicators
stretching lineation
and shear sense
(thrusting and exhumation)

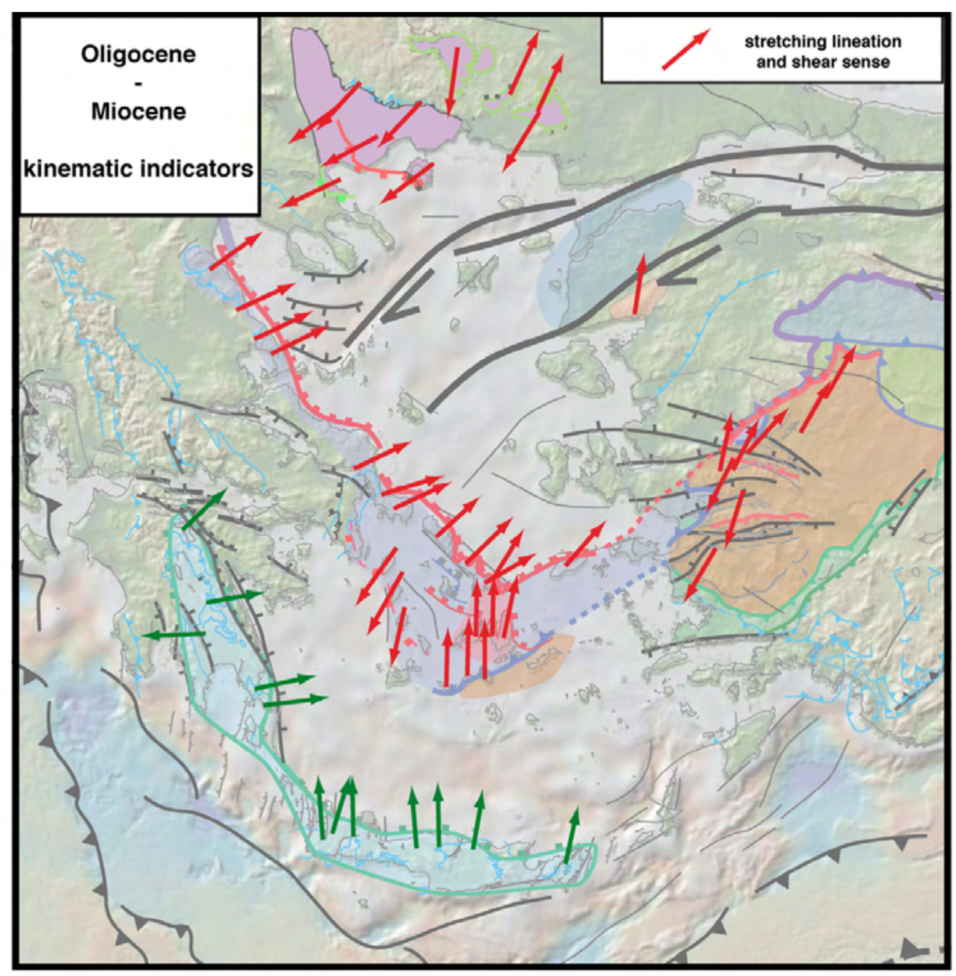

Oligocene
-
Miocene
kinematic indicators
stretching lineation
and shear sense

Figure 5

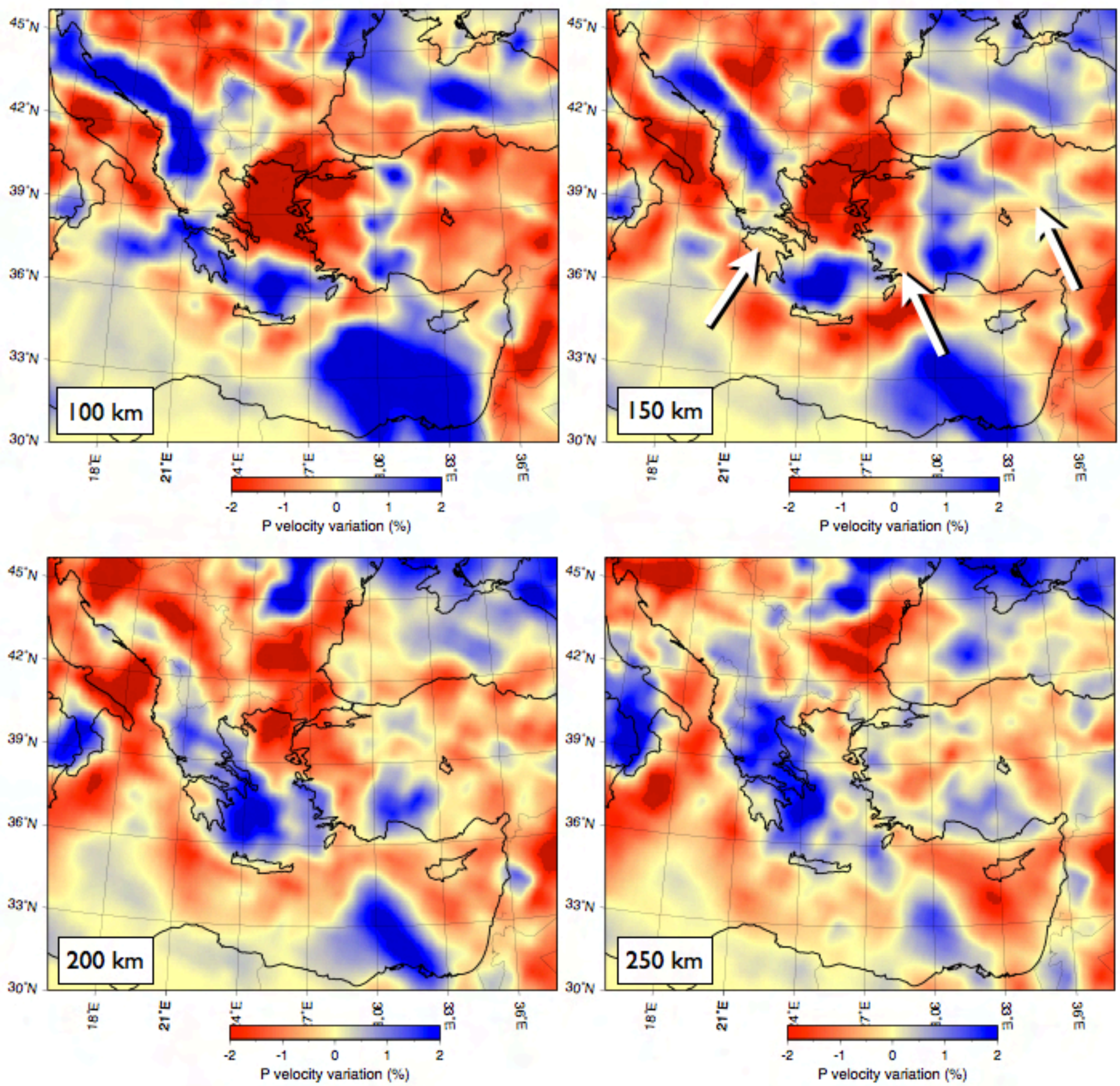

100 km
150 km
200 km
250 km
45°N
42°N
39°N
36°N
33°N
30°N
18°E
21°E
P velocity variation (%)
-2
-1
0
1
2

Figure 6

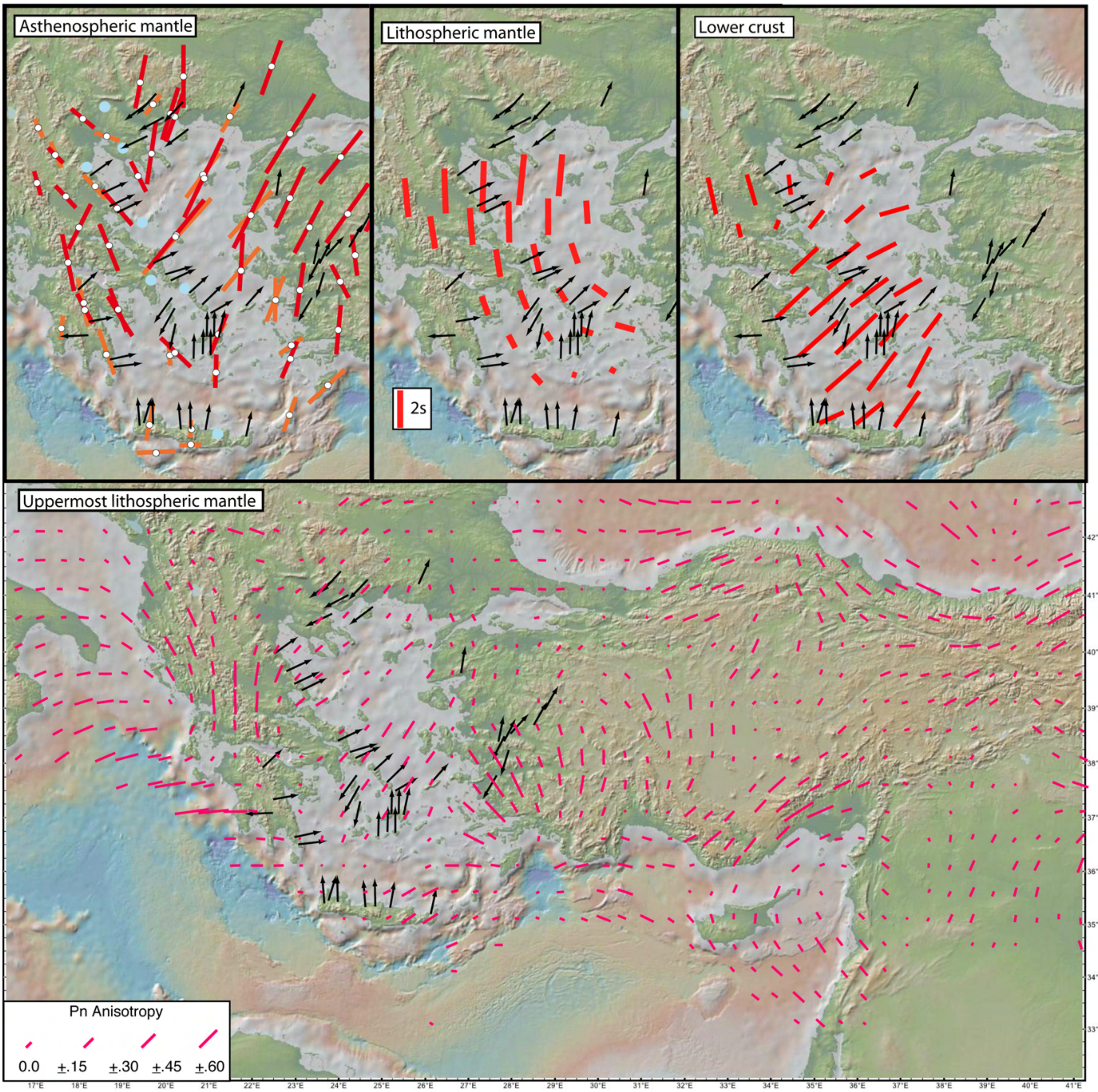
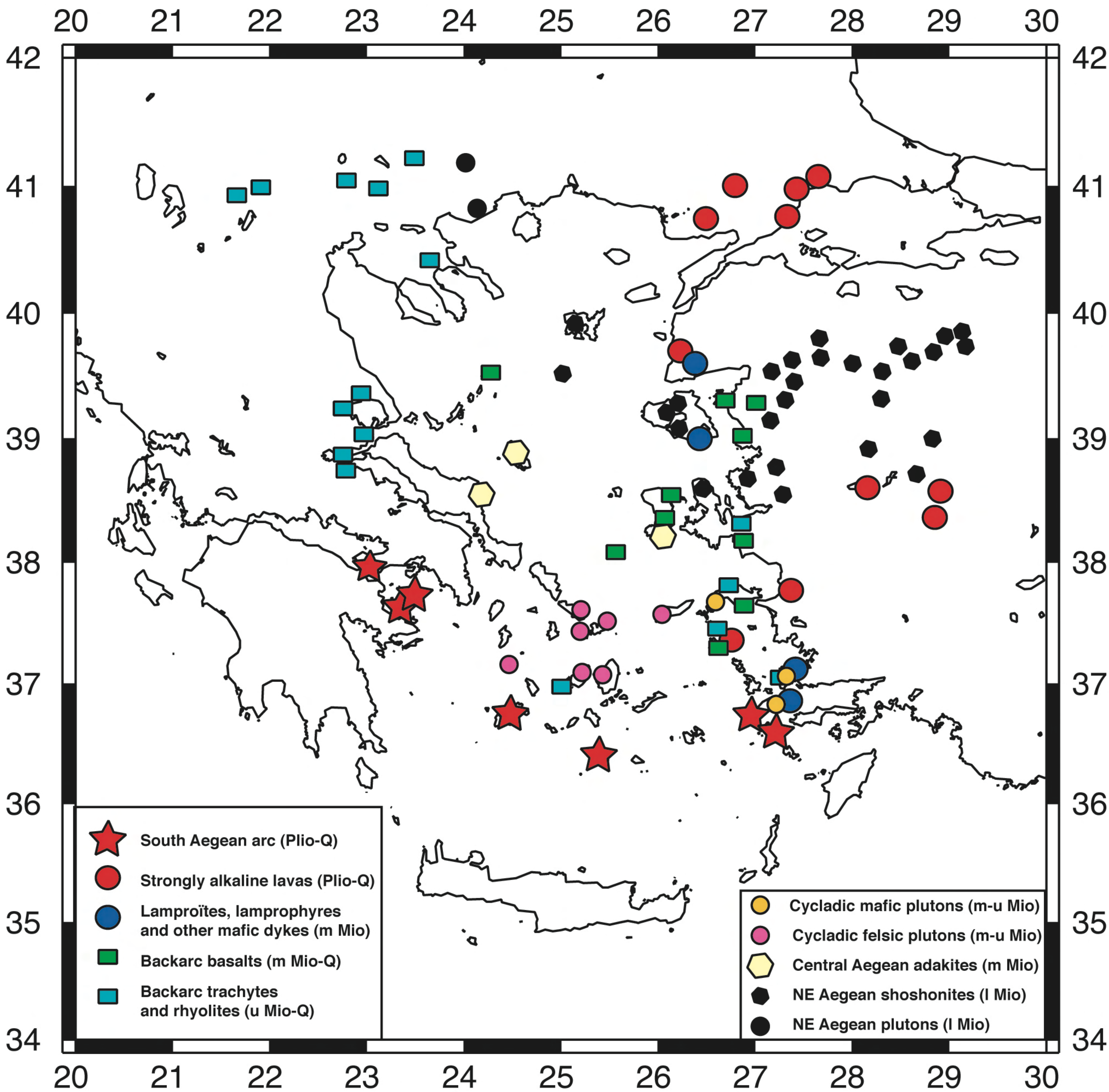
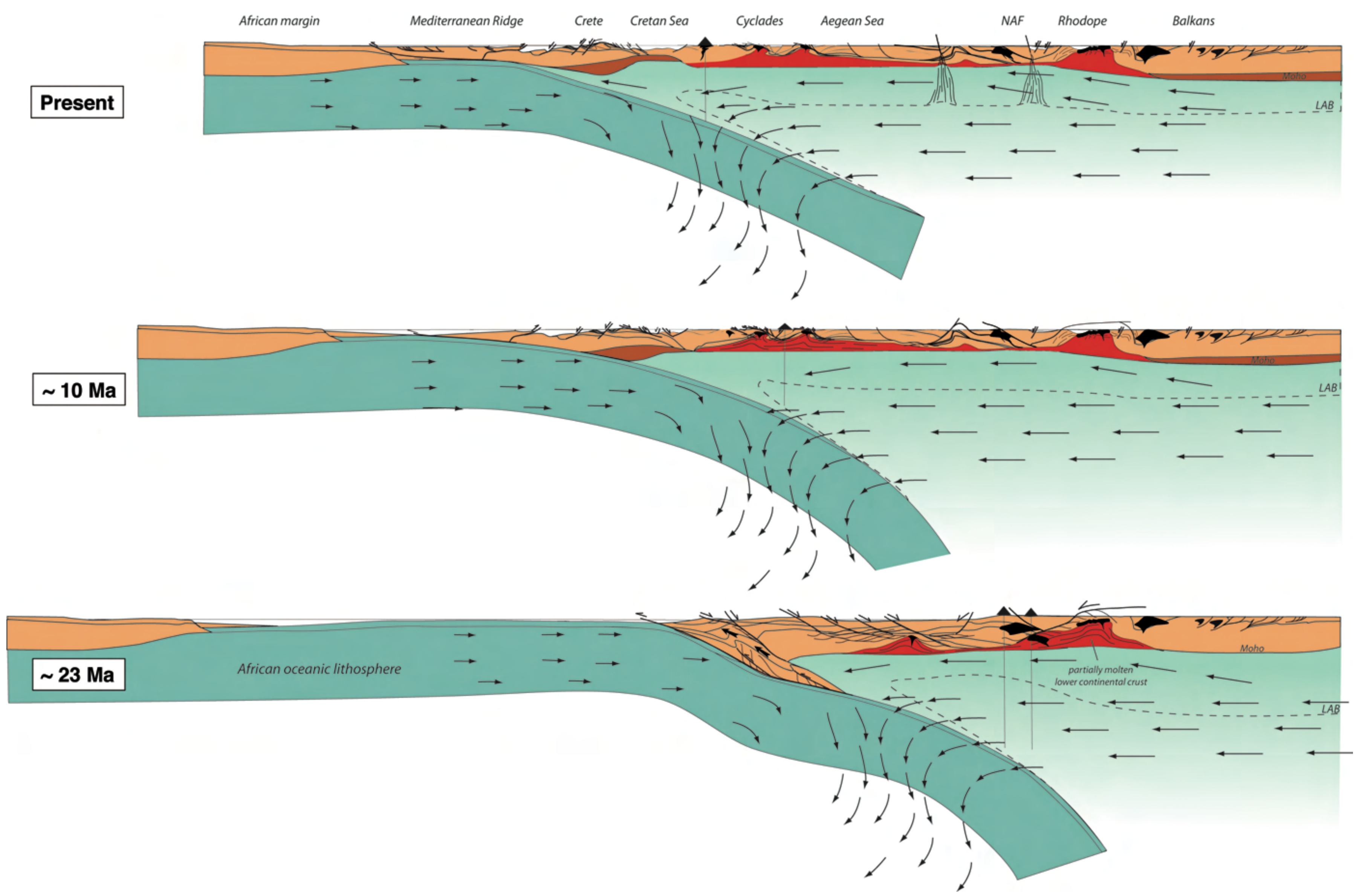

Asthenospheric mantle
Lithospheric mantle
2s
Lower crust
Uppermost lithospheric mantle
Pn Anisotropy
0.0
±.15
±.30
±.45
±.60

Figure 7



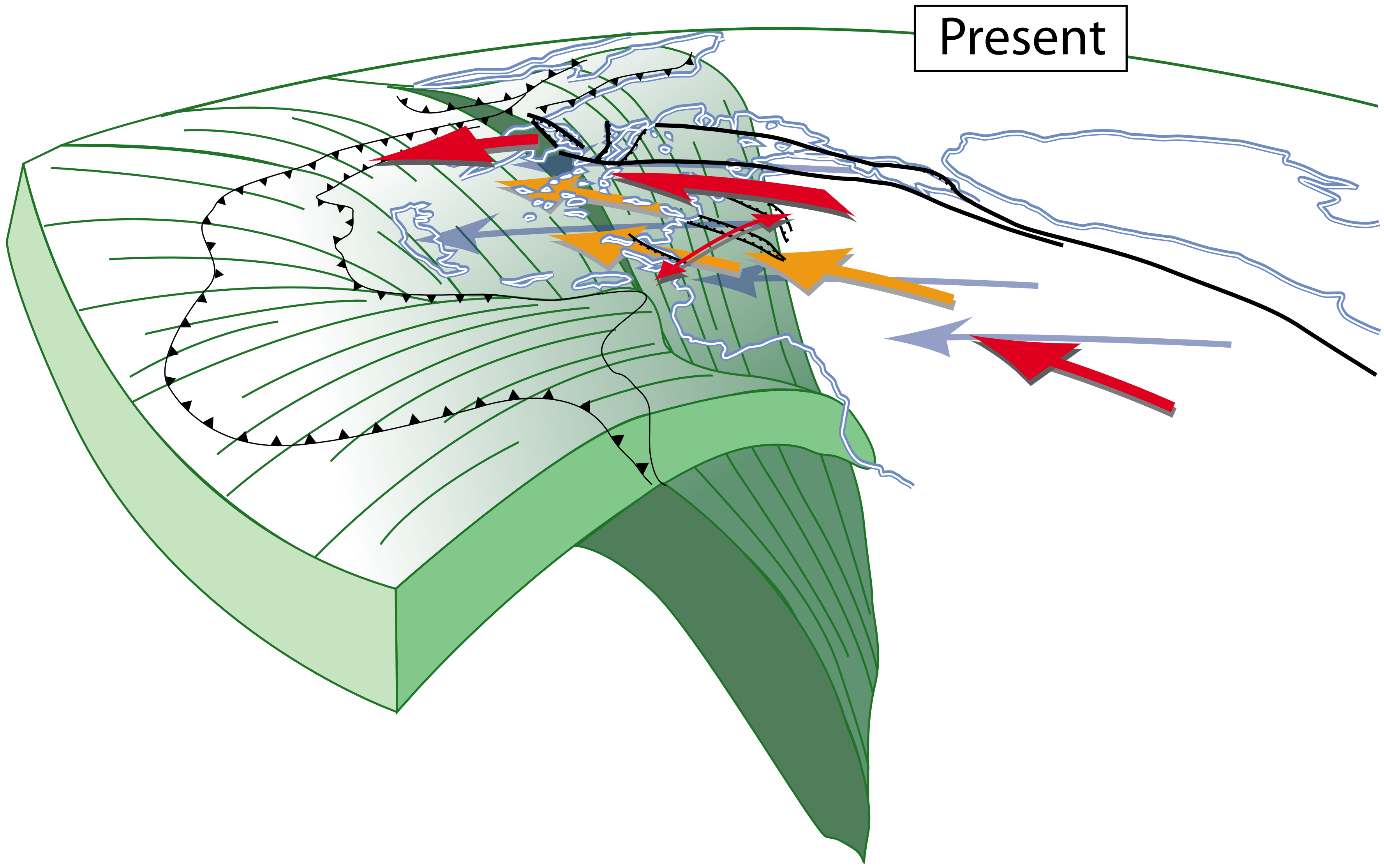
20
21
22
23
24
25
26
27
28
29
30
42
41
40
39
38
37
36
35
34
South Aegean arc (Plio-Q)
Strongly alkaline lavas (Plio-Q)
Lamproïtes, lamprophyres and other mafic dykes (m Mio)
Backarc basalts (m Mio-Q)
Backarc trachytes and rhyolites (u Mio-Q)
Cycladic mafic plutons (m-u Mio)
Cycladic felsic plutons (m-u Mio)
Central Aegean adakites (m Mio)
NE Aegean shoshonites (l Mio)
NE Aegean plutons (l Mio)

Figure 8



African margin
Mediterranean Ridge
Crete
Cretan Sea
Cyclades
Aegean Sea
NAF
Rhodope
Balkans
Present
Moho
LAB
~ 10 Ma
Moho
LAB
~ 23 Ma
African oceanic lithosphere
partially molten lower continental crust
Moho
LAB

Figure 9



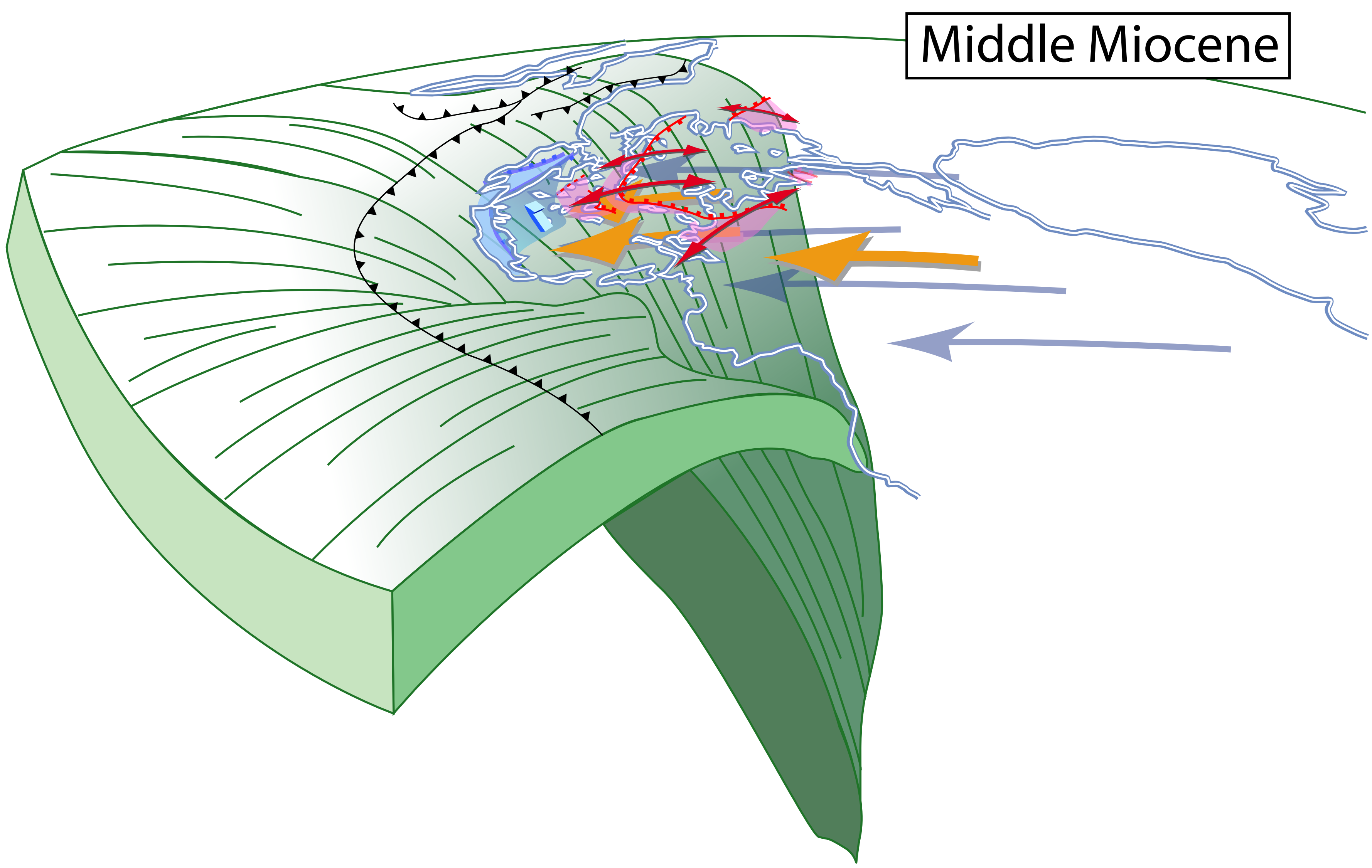


Figure 10

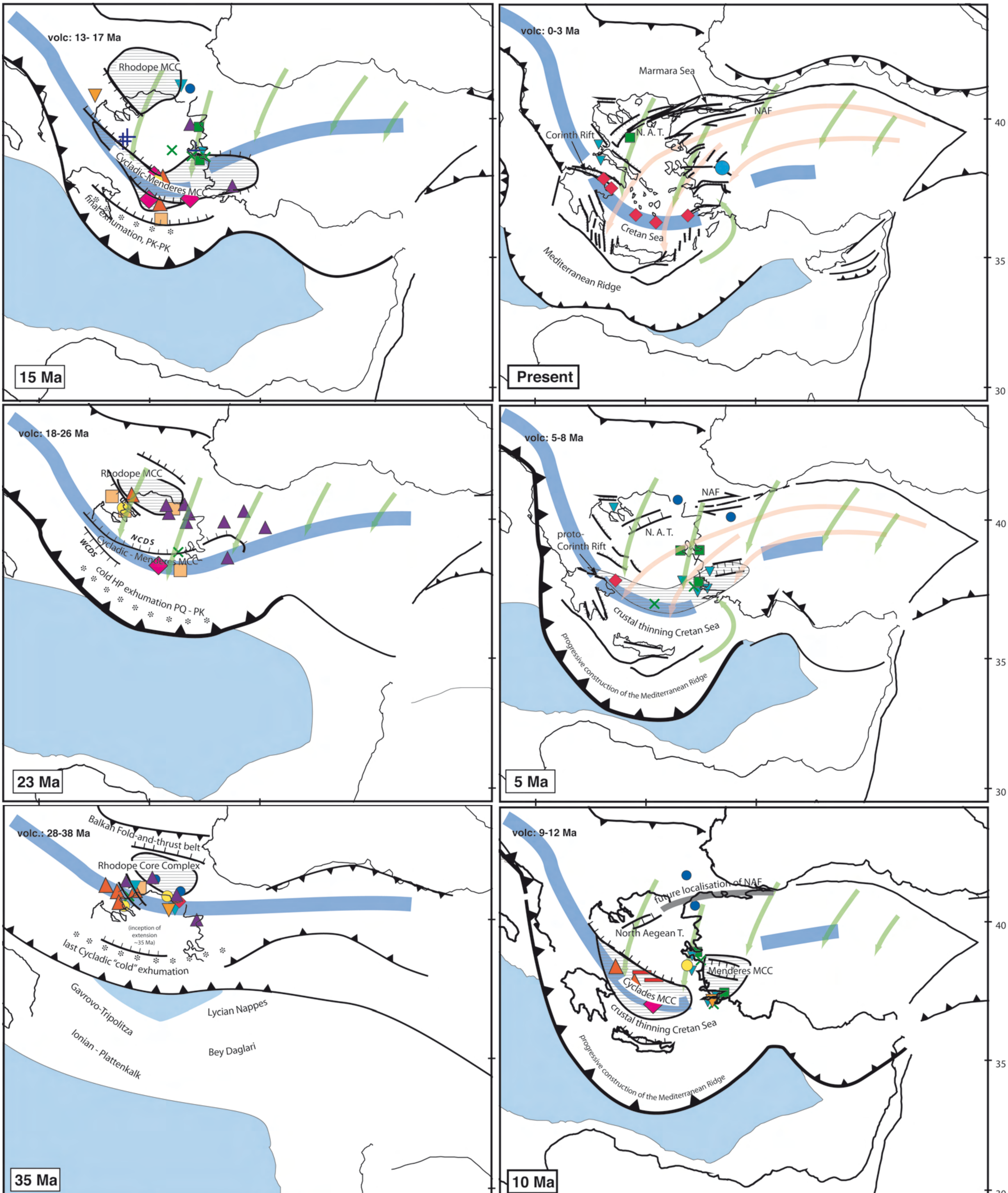

volc: 13- 17 Ma
Rhodope MCC
Cycladic-Menderes MC
final exhumation, PK-PK
15 Ma
volc: 0-3 Ma
Marmara Sea
NAF
Corinth Rift
N. A. T.
Cretan Sea
Mediterranean Ridge
Present
volc: 18-26 Ma
Rhodope MCC
NCDS
WCDS
Cycladic - Menderes MCC
cold HP exhumation PQ - PK
23 Ma
volc: 5-8 Ma
NAF
proto-
Corinth Rift
N. A. T.
crustal thinning Cretan Sea
progressive construction of the Mediterranean Ridge
5 Ma
volc.: 28-38 Ma
Balkan Fold-and-thrust belt
Rhodope Core Complex
(inception of
extension
~35 Ma)
last Cycladic "cold" exhumation
Gavrovo-Tripolitza
Ionian - Plattenkalk
Lycian Nappes
Bey Daglari
35 Ma
volc: 9-12 Ma
future localisation of NAF
North Aegean T.
Menderes MCC
Cyclades MCC
crustal thinning Cretan Sea
progressive construction of the Mediterranean Ridge
10 Ma
40 N
35 N
30 N

Figure 11